
CLM5 Documentation

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This document has two major sections.

CHAPTER 1

CLM USER'S GUIDE

1.1 Overview

User's Guide to version 5.0 of the Community Land Model (CLM)

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1.1.1 Introduction

The Community Land Model (CLM5.0 in CESM2.0) is the latest in a series of global land models developed by the CESM Land Model Working Group (LMWG) and maintained at the National Center for Atmospheric Research (NCAR). This guide is intended to instruct both the novice and experienced user on running CLM. This guide pertains to the latest version CLM5.0 in CESM2.0 available for download from the public release subversion repository as a part of CESM1.2.0. Documentation may be different if you are using an older version, you should either use the documentation for that release version, update to the latest version, or use the documentation inside your own source tree. There is information in the ChangeLog file and in the [What is new with CLM5.0 in CESM2.0 since previous public releases?](#) regarding the changes from previous versions of CESM.

Note: This release of CLM5.0 in CESM1.2.0 includes BOTH CLM4.0 physics used in previous releases as well as the updated CLM4.5 physics. Both CLM as well as CLM support tools allow you to trigger between the two physics modes. Most often when we refer to CLM4.0 we are referring to the CLM4.0 physics in CLM4.5 in CESM1.2.0 rather than to a specific version of CLM4.0 (where we would give the exact version). Likewise, when referring to CLM4.5 we are referring to the CLM4.5 physics in CLM4.5 in CESM1.2.0.

The novice user should read [Chapter 1](#) in detail before beginning work, while the expert user should read [What is new with CLM4.5 in CESM1.2.0 since previous public releases?](#) and [Quickstart to using CLM4.5](#) chapters, and then use

the more detailed chapters as reference. Before novice users go onto more technical problems covered in [Chapter 2](#), [Chapter 3](#), [Chapter 4](#), or [Chapter 5](#) they should know the material covered in [Chapter 1](#) and be able to replicate some of the examples given there.

All users should read the [How to Use This Document](#) and [Other resources](#) to get help from sections to understand the document conventions and the various ways of getting help on using CLM4.5. Users should also read the [What is scientifically validated and functional in CLM4.5 in CESM1.2.0?](#) section to see if their planned use of the model is something that has been scientifically validated and well tested. Users that are NOT using NCAR machines or our list of well tested machines should also read the [What are the UNIX utilities required to use CLM4.5?](#) section to make sure they have all the required UNIX utilities on the system they want to do their work.

Developers that are making changes to CLM either for their own development or for development that they hope will eventually become a part of the main CLM should read the [Chapter 8](#) chapter. We have a suite of test scripts that automatically test many different model configurations and namelist options, as well as ensuring things like restarts are bit-for-bit and the like. It's helpful to use these scripts to ensure your changes are working correctly. As well as being a required part of the process to bring in new code developments. And it's far easier to use the automated scripts rather than having to figure out, what to test, how to do it, and then finally do it by hand. If you are using non supported machines you may also want to use the test scripts to make sure your machine is working correctly.

1.1.2 What is New with CLM4.5

The CESM1.2.0 [What's New Science](#) and [What's New Software](#) pages gives a synopsis of the changes to all CESM components since the CESM1.1.1 release. More details are given in the [CLM ChangeLog file](#).

Previous release pages give similar list of changes for previous versions of the model. The [CLM4 in CESM1.0.5 User's Guide](#) gives information on the updates for versions up to CLM4 in CESM1.0.5.

1.1.3 Overview of User's Guide

In this introduction we first give a simple guide to understand the document conventions in [How to Use This Document](#). The next section [What is new with CLM4.5 in CESM1.2.0 since previous public releases?](#) gives references to describe the differences between CLM4.5 in CESM1.2.0 and previous CESM releases both from a scientific as well as a software engineering point of view. For information on previous releases of CLM4.5 before CLM4.5 in CESM1.2.0 see the CESM1.1.1 documentation. The next section [Quickstart to using CLM4.5](#) is for users that are already experts in using CLM and gives a quickstart guide to the bare details on how to use CLM4.5. The next [What is scientifically validated and functional in CLM4.5 in CESM1.2.0?](#) tells you about what has been extensively tested and scientifically validated (and maybe more importantly) what has NOT. [What are the UNIX utilities required to use CLM4.5?](#) lists the UNIX utilities required to use CLM4.5 and is important if you are running on non-NCAR machines, generic local machines, or machines NOT as well tested by us at NCAR. Next we have [Important Notes and Best Practices for Usage of CLM4.5](#) to detail some of the best practices for using CLM4.5 for science. The last introductory section is [Other resources](#) to get help from which lists different resources for getting help with CESM1.0 and CLM4.5.

[Chapter 1](#) goes into detail on how to setup and run simulations with CLM4.5 and especially how to customize cases. Details of cesm_setup modes and build-namelist options as well as namelist options are given in this chapter.

[Chapter 2](#) gives instructions on the CLM tools for either CLM4.0 or CLM4.5 physics for creating input datasets for use by CLM, for the expert user. There's an overview of what each tool does, and some general notes on how to build the FORTRAN tools. Then each tool is described in detail along with different ways in which the tool might be used. A final section on how to customize datasets for observational sites for very savvy expert users is given as the last section of this chapter.

As a followup to the tools chapter, [Chapter 3](#) tells how to add files to the XML database for build-namelist to use. This is important if you want to use the XML database to automatically select user-created input files that you have created when you setup new cases with CLM (both CLM4.0 and CLM4.5 physics).

In Chapter 4, again for the expert user, we give details on how to do some particularly difficult special cases. For example, we give the protocol for spinning up the CLM4.5-BGC and CLMCN models as well as CLM with dynamic vegetation active (CNDV). We give instructions to do a spinup case from a previous case with Coupler history output for atmospheric forcing. We also give instructions on running both the prognostic crop and irrigation models. Lastly we tell the user how to use the DATM model to send historical CO₂ data to CLM.

Chapter 5 outlines how to do single-point or regional simulations using CLM4.5. This is useful to either compare CLM4.5 simulations with point observational stations, such as tower sites (which might include your own atmospheric forcing), or to do quick simulations with CLM for example to test a new parameterization. There are several different ways given on how to perform single-point simulations which range from simple PTS_MODE to more complex where you create all your own datasets, tying into Chapter 2 and also Chapter 3 to add the files into the build-namelist XML database. The PTCLM python script to run single-point simulations was added back in for this release (but it has bugs that don't allow it to work out of the box). CLM4 in CESM1.0.5 has a fully working versions of PTCLM.

Need Chapter 6 blurb...

Chapter 7 gives some guidance on trouble-shooting problems when using CLM4.5. It doesn't cover all possible problems with CLM, but gives you some guidelines for things that can be done for some common problems.

Chapter 8 goes over the automated testing scripts for validating that the CLM is working correctly. The test scripts run many different configurations and options with CLM4.0 physics as well and CLM4.5 physics making sure that they work, as well as doing automated testing to verify restarts are working correctly, and testing at many different resolutions. In general this is an activity important only for a developer of CLM4.5, but could also be used by users who are doing extensive code modifications and want to ensure that the model continues to work correctly.

In the appendices we talk about some issues that are useful for advanced users and developers of CLM4.5.

Finally in Appendix A we give instructions on how to build the documentation associated with CLM4.5 (i.e. how to build this document). This document is included in every CLM distribution and can be built so that you can view a local copy rather than having to go to the CESM website. This also could be useful for developers who need to update the documentation due to changes they have made.

1.1.4 Best Practices

- CLM4.5 includes BOTH the old CLM4.0 physics AND the new CLM4.5 physics and you can toggle between two. The “standard” practice for CLM4.0 is to run with CN on, and with Qian atmospheric forcing. While the “standard” practice for CLM4.5 is to run with BGC on, and CRUNCEP atmospheric forcing. “BGC” is the new CLM4.5 biogeochemistry and include CENTURY-like pools, vertical resolved carbon, as well as Nitrification and de-Nitrification (see the Section called Some Acronym's and Terms We'll be Using in Other resources to get help from).
- When running with CLMCN (either CLM4.0 or CLM4.5 physics) or CLM4.5-BGC, it is critical to begin with initial conditions that are provided with the release or to spin the model up following the CN spinup procedure before conducting scientific runs (see the Section called Spinning up the CLM4.5 biogeochemistry (CLMBGC spinup) in Chapter 4 for CLM4.5 or the Section called Spinning up the CLM4.0 biogeochemistry Carbon-Nitrogen Model (CN spinup) in Chapter 4 for CLM4.0). Simulations without a proper spinup will effectively be starting from an unvegetated world. See the Section called Setting Your Initial Conditions File in Chapter 1 for information on how to provide initial conditions for your simulation.
- Initial condition files are provided for CLM4.0-CN as before, for fully coupled BCN and offline ICN cases for 1850 and 2000 at finite volume grids: 1deg (0.9x1.25), 2deg (1.9x2.5), and T31 resolutions. We also have interpolated initial conditions for BCN for 1850 and 2000 for two finite volume grids: 10x15, 4x5 and two HOMME grids (ne30np4 and ne120np4). There's also an initial condition file for ICN with the prognostic crop model for 2000 at 2deg resolution, and one with CLMSP for 2000 at 2deg resolution. We also have initial conditions for offline CNDV for 1850. The 1850 initial condition files are in ‘reasonable’ equilibrium. The 2000 initial condition files represent the model state for the year 2000, and have been taken from transient simulations. Therefore, by design the year 2000 initial condition files do not represent an equilibrium state.

Note also that spinning the 2000 initial conditions out to equilibrium will not reflect the best estimate of the real carbon/nitrogen state for the year 2000.

- Initial condition files are also provided for CLM4.5 for several configurations and resolutions. For CLM4.5-SP and CLM4.5-BGC with CRUNCEP forcing we have initial conditions at 1deg resolution for both 1850 and 2000. The CLM4.5-BGC initial conditions for 1850 (again with CRUNCEP forcing) were also interpolated to 2deg, CRUNCEP half degree (360x720cru), and ne30np4 resolutions. Also the CLM4.5-BGC initial conditions for 1850 (with CRUNCEP forcing) were interpolated to 1deg CLM4.5-BGC-DV and 2deg CLM4.5-BGC-Crop.
- Users can generate initial condition files at different resolutions by using the CLM tool interpinic to interpolate from one of the provided resolutions to the resolution of interest. Interpolated initial condition files may no longer be in ‘reasonable’ equilibrium.
- In CLM4.5 for both CLM4.5-CN and CLM4.5-BGC the new fire model requires lightning frequency data, and human population density (both are read inside of CLM). By default we have provided a climatology dataset for lightning frequency and a dataset with coverage from 1850 to 2010 for population density. Both of these datasets are interpolated from the native resolution of the datasets to the resolution you are running the model on. If you are running with an atmosphere model or forcing that is significantly different than present day – the lightning frequency may NOT appropriately correspond to your atmosphere forcing and fire initiation would be inappropriate.
- Aerosol deposition is a required field to both CLM4.0 and CLM4.5 physics, sent from the atmosphere model. Simulations without aerosol deposition will exhibit unreasonably high snow albedos. The model sends aerosol deposition from the atmospheric model (either CAM or DATM). When running with prescribed aerosol the atmosphere model will interpolate the aerosols from 2-degree resolution to the resolution the atmosphere model is running at.

1.1.5 How To Use This Document

Links to descriptions and definitions have been provided in the code below. We use the same conventions used in the CESM documentation as outlined below.

Throughout the document this style is used to indicate shell commands and options, fragments of code, namelist variables, etc. Where examples from an interactive shell session are presented, lines starting with > indicate the shell prompt. A backslash \" at the end of a line means the line continues onto the next one (as it does in standard UNIX shell). Note that \$EDITOR" is used to refer to the text editor of your choice. \$EDITOR is a standard UNIX environment variable and should be set on most UNIX systems. Comment lines are signaled with a "#" sign, which is the standard UNIX comment sign as well. \$CSMDATA is used to denote the path to the inputdata directory for your CESM data.

```
> This is a shell prompt with commands \
that continues to the following line.
> $EDITOR filename # means you are using a text editor to edit "filename"
# This is a comment line
```

1.1.6 Quickstart

Running the CLM requires a suite of UNIX utilities and programs and you should make sure you have all of these available before trying to go forward with using it. If you are missing one of these you should contact the systems administrator for the machine you wish to run on and make sure they are installed.

List of utilities required for CESM in the “CESM1.2.0 Software/Operating System Prerequisites” section in <http://www.cesm.ucar.edu/models/cesm1.2//cesm/doc/usersguide/book1.html> - UNIX bash shell (for some of the CLM tools scripts) - NCL (for some of the offline tools for creating/modifying CLM input datasets see [Chapter 2](#) for more information on NCL) - Python (optional, needed for PTCLM) - xsltproc, docbook and docbook utilities (optional, needed to build the Users-Guide)

Before working with CLM4.5 read the QuickStart Guide in the [CESM1.2.0 Scripts User’s Guide](#). Once you are familiar with how to setup cases for any type of simulation with CESM you will want to direct your attention to the specifics of using CLM.

For some of the details of setting up cases for CLM4.5 read the README and text files available from the “models/lnd/clm/doc” directory (see the “CLM Web pages” section for a link to the list of these files). Here are the important ones that you should be familiar with.

1. [README file](#) describing the directory structure.
2. [Quickstart.userdatasets](#) file describing how to use your own datasets in the model (also see the Section called [Creating your own single-point/regional surface datasets in Chapter 5](#)).
3. [models/lnd/clm/doc/KnownBugs](#) file describing known problems in CLM4.5 (that we expect to eventually fix).
4. [models/lnd/clm/doc/KnownLimitations](#) file describing known limitations in CLM4.5 and workarounds that we do NOT expect to fix.

The IMPORTANT_NOTES file talks about important things for users to know about using the model scientifically. Its content is given in the next chapter on “[What is scientifically validated and functional in CLM4.5 in CESM1.2.0?](#)”.

The ChangeLog/ChangeSum talk about advances in different versions of CLM. The content of these files is largely explained in the previous chapter on “[What is new with CLM4.5 in CESM1.2.0 since previous public releases?](#)”.

Note other directories have README files that explain different components and tools used when running CLM and are useful in understanding how those parts of the model work and should be consulted when using tools in those directories. For more details on configuring and customizing a case with CLM see [Chapter 1](#).

The Quickstart.GUIDE (which can be found in models/lnd/clm/doc) is repeated here.

```
Quick-Start to Using cpl7 Scripts for clm4_5

Assumptions: You want to use yellowstone with clm4_5 BGC
              to do a clm simulation with data atmosphere and the
              latest CRUNCEP atm forcing files and settings. You also want to cycle
              the CRUNCEP atm data between 1901 to 1920 and you want to run at
              0.9x1.25 degree resolution.

Process:

# Create the case

cd scripts

./create_newcase -case <testcase> -mach yellowstone_intel -res f09_g16 -
↪compset I1850CRUCLM45BGC
(./create_newcase -help -- to get help on the script)

# Setup the case

cd <testcase>
./xmlchange id1=val1,id2=val2 # to make changes to any settings in the env_*.xml
↪files
./cesm_setup
(./cesm_setup -help -- to get help on the script, this creates the ./<testcase>.run \
```

```
script)

# Add any namelist changes to the user_nl_* files

$EDITOR user_nl_*

# Compile the code

./<testcase>.build

# Submit the run

./<testcase>.submit
```

Information on Compsets:

“I” compsets are the ones with clm and datm7 without ice and ocean. They specify either CLM4.0 physics or CLM4.5 physics. Most of the “I” compsets for CLM4.0 use the CLM_QIAN data with solar following the cosine of solar zenith angle, precipitation constant, and other variables linear interpolated in time (and with appropriate time-stamps on the date). Useful “I” compsets for CLM4.5 use the CRUNCEP data in place of the CLM_QIAN data.

To list all the compsets use: ./create_newcase -list compsets

Some of the CLM4.5 I compsets are:

Alias Description 1850CRUCLM45 CLM4.5 to simulate year=1850 with CLMN45SP (Satellite Phenology)
I1850CRUCLM45BGC CLM4.5 to simulate year=1850 with CLM45BGC biogeophysics model (BGC)
I20TRCRUCLM45BGC CLM4.5 with BGC on with transient PFT over 1850-2000

While some of the CLM4 I compsets are:

Alias Description ICN CLM4.0 to simulate year=2000 with Carbon-Nitrogen BGC model (CN) I1850CN
CLM4.0 to simulate year=1850 with Carbon-Nitrogen BGC model (CN) I20TRCN CLM4.0 with CN on with transient PFT over 1850-2000
IRCP26CN CLM4.0 with CN on with transient PFT over 1850-2100 for RCP=2.6 scenario
IRCP45CN CLM4.0 with CN on with transient PFT over 1850-2100 for RCP=4.5 scenario
IRCP60CN CLM4.0 with CN on with transient PFT over 1850-2100 for RCP=6.0 scenario
IRCP85CN CLM4.0 with CN on with transient PFT over 1850-2100 for RCP=8.5 scenario

Automatically resubmitting jobs:

After doing a short simulation that you believe is correct

```
./xmlchange CONTINUE_RUN=TRUE  
# Change RESUBMIT to number greater than 0, and CONTINUE_RUN to TRUE...  
./<testcase>.submit
```

1.1.7 Scientific Validation

In this section we go over what has been extensively tested and scientifically validated with CLM4.5, and maybe more importantly what has NOT been tested and may NOT be scientifically validated. You can use all features of CLM, but need to realize that some things haven’t been tested extensively or validated scientifically. When you use these features you may run into trouble doing so, and will need to do your own work to make sure the science is reasonable.

Standard Configuration and Namelist Options that are Validated

See http://www.cesm.ucar.edu/models/cesm1.2/clm/CLM_configurations_CESM1.2.pdf for an explanation of what configurations are scientifically validated for CLM4.5. For CLM4.0 changes to the science of the model are minimal since CESM1.1.1 so we expect answers to be very similar to using it.

In the sections below we go through configuration and/or namelist options or modes that the user should be especially wary of using. You are of course free to use these options, and you may find that they work functionally. Although in some cases you will find issues even with functionality of using them. If so you will need to test, debug and find solutions for these issues on your own. But in every case you will need to go through more extensive work to validate these options from a scientific standpoint. Some of these options are only for CLM4.5 while others are for both CLM4.0 AND CLM4.5 we explicitly say which they apply to.

Configure Modes NOT scientifically validated, documented, supported or, in some cases, even advised to be used:

These are options that you would add to `CLM_CONFIG_OPTS`.

1. exlaklayers on[CLM4.5 only] This mode is NOT tested and may NOT be even functional.
2. snicar_frc on[CLM4.0 AND CLM4.5] This mode is tested and functional, but is NOT constantly scientifically validated, and should be considered experimental.
3. vichydro on[CLM4.5 only] This mode is tested and functional, but does NOT have long scientific validation simulations run with it so, should be considered experimental.
4. vsoilc_centbgc[CLM4.5 only] This option is extensively tested for both “on” and “off”. The “no-vert” option has limited testing performed on it, but isn’t scientifically validated (and it currently has a bug – see 1746 and 1672 in [models/lnd/clm/doc/KnownBugs](#)). The “no-cent” and “no-nitrif” options are NOT tested and as such may NOT be even functional.

Namelist options that should NOT be exercised:

Build-Namelist options that should NOT be exercised:

1. -irrig with -bgc cn and -phys clm4_0 We have only run the irrigation model with CLMSP (i.e. without the CN model). We recommend that if you want to run the irrigation model with CN, that you do a spinup. But, more than that you may need to make adjustments to irrig_factor in [models/lnd/clm/src/biogeophys/CanopyFluxesMod.F90](#). See the notes on this in the description of the irrigation model in the [Technical Descriptions of the Interactive Crop Management and Interactive Irrigation Models](#).
2. -irrig with -crop on and -phys clm4_0 Irrigation doesn’t work with the prognostic crop model. Irrigation is only applied to generic crop currently, which negates its practical usage. We also have a known problem when both are on (see bug 1326 in the [models/lnd/clm/doc/KnownBugs](#) file). If you try to run in this mode, the CLM build-namelist will return with an error.

Namelist items that should NOT be exercised:

`suplnitro='ALL'` The suplnitro namelist option to the CN Biogeochemistry model supplies unlimited nitrogen and therefore vegetation is over-productive in this mode.

`urban_traffic`:Not currently functional

`allowlakeprod`:Considered experimental.

`anoxia_wtsat`:Considered experimental (deprecated will be removed).

atm_c14_filename:Considered experimental (dataset not provided).
exponential_rooting_profile:Considered experimental.
fin_use_fsat:Considered experimental.
glc_dyntopo:Not currently functional.
lake_decomp_fact:Considered experimental.
more_vertlayers:Considered experimental.
no_frozen_nitrif_denitrif:Considered experimental.
perchroot:Considered experimental.
perchroot_alt:Considered experimental.
replenishlakec:Considered experimental.
use_c14_bombspike:Considered experimental (dataset not provided).
usefrootc:Considered experimental.

1.1.8 Getting Help

In addition to this users-guide there are several other resources that are available to help you use CLM5.0. The first one is the CESM1.2.0 User's-Guide, which documents the entire process of creating cases with CESM1.2.0. The next is the CESM bulletin board which is a web-site for exchanging information between users of CESM. There are also CLM web-pages specific for CLM, and finally there is an email address to report bugs that you find in CESM1.2.0.

The CESM User's-Guide

CLM5.0 in CESM2.0 is always run from within the standard CESM2.0 build and run scripts. Therefore, the user of CLM4.5 should familiarize themselves with the CESM1.2.0 scripts and understand how to work with them. User's-Guide documentation on the CESM1.2.0 scripts are available from the following web-page. The purpose of this CLM4.5 in CESM1.2.0 User's Guide is to give the CLM4.5 user more complete details on how to work with CLM and the set of tools that support CLM, as well as to give examples that are unique to the use of CLM. However, the CESM1.2.0 Scripts User's-Guide remains the primary source to get detailed information on how to build and run the CESM system.

cesmrel; Scripts User's-Guide (<http://www.cesm.ucar.edu/models/cesm1.2/cesm/doc/usersguide/book1.html>)

The CESM Bulletin Board

There is a rich and diverse set of people that use the CESM, and often it is useful to be in contact with others to get help in solving problems or trying something new. To facilitate this we have an online Bulletin Board for questions on the CESM. There are also different sections in the Bulletin Board for the different component models or for different topics.

CESM Online Bulletin Board

The CLM web pages

The main CLM web page contains information on the CLM, its history, developers, as well as downloads for previous model versions. There are also documentation text files in the models/lnd/clm/doc directory that give some quick information on using CLM.

CLM web page CLM Documentation Text Files

Also note that several of the XML database files can be viewed in a web browser to get a nice table of namelist options, namelist defaults, or compsets. Simply view them as a local file and bring up one of the following files:

- `models/lnd/clm/bld/namelist_files/namelist_definition_clm4_0.xml` – definition of CLM4.0 namelist items.
- `models/lnd/clm/bld/namelist_files/namelist_definition_clm4_5.xml` – definition of CLM4.0 namelist items.
- `models/lnd/clm/bld/namelist_files/namelist_defaults_clm4_0.xml` – default values for CLM4.0 namelist items.
- `models/lnd/clm/bld/namelist_files/namelist_defaults_clm4_5.xml` – default values for CLM4.5 namelist items.
- `scripts/ccsm_utils/Case.template/config_definition.xml` – definition of all env_*.xml items.
- `scripts/ccsm_utils/Case.template/config_compsets.xml` – definition of all the compsets.
- `models/lnd/clm/bld/namelist_files/history_fields_clm4_0.xml` – definition of CLM4.0 history fields.
- `models/lnd/clm/bld/namelist_files/history_fields_clm4_5.xml` – definition of CLM4.5 history fields.

Reporting bugs in CLM4.5

If you have any problems, additional questions, bug reports, or any other feedback, please send an email to <cesmhelp@cgd.ucar.edu>. If you find bad, wrong, or misleading information in this users guide send an email to <erik@ucar.edu>. The current list of known issues for CLM4.5 in CESM1.2.0 is in the `models/lnd/clm/doc/KnownBugs` file, and the list of issues for CESM1.2.0 is at... http://www.cesm.ucar.edu/models/cesm1.2//tags/cesm1_2_0/#PROBLEMS.

Some Acronym's and Terms We'll be Using

CAM Community Atmosphere Model (CAM). The prognostically active atmosphere model component of CESM.

CESM Community Earth System Model (CESM). The coupled earth system model that CLM is a component of.

CLM Community Land Model (CLM). The prognostically active land model component of CESM.

CLMBGC Community Land Model (CLM4.5) with BGC Biogeochemistry. Uses CN Biogeochemistry with vertically resolved soil Carbon, CENTURY model like pools, and Nitrification/De-Nitrification. The `CLM_CONFIG_OPTS` option for this is

```
./xmlchange CLM_CONFIG_OPTS="phys clm4_5 -bgc cn -vsoilc_centbgc on
-clm4me on"
```

CLMCN Community Land Model (CLM) with Carbon Nitrogen (CN) Biogeochemistry (either CLM4.0 or CLM4.5)

The `CLM_CONFIG_OPTS` option for this is

```
./xmlchange CLM_CONFIG_OPTS="-bgc cn" -append
```

CLMSP Community Land Model (CLM) with Satellite Phenology (SP) (either CLM4.0 or CLM4.5) The `CLM_CONFIG_OPTS` option for this is

```
./xmlchange CLM_CONFIG_OPTS="-bgc none" -append
```

CLMU Community Land Model (CLM) Urban Model (either CLM4.0 or CLM4.5). The urban model component of CLM is ALWAYS active (unless you create special surface datasets that have zero urban percent, or for regional/single-point simulations for a non-urban area).

CRUNCEP The Climate Research Unit (CRU) analysis of the NCEP atmosphere reanalysis atmosphere forcing data. This can be used to drive CLM with atmosphere forcing from 1901 to 2010. We also DO expect to be able to update this dataset beyond 2010 as newer data becomes available.

DATM Data Atmosphere Model (DATM) the prescribed data atmosphere component for CESM. Forcing data that we provide are either the Qian or CRUNCEP forcing datasets (see below).

DV Dynamic global vegetation, where fractional PFT (see PFT below) changes in time prognostically. Can NOT be used with prescribed transient PFT (requires either CLM4.5-BGC or CLMCN for either CLM4.0 or CLM4.5). The CLM_CONFIG_OPTS option for this is

```
./xmlchange CLM_CONFIG_OPTS="-bgc cndv" -append
```

ESMF Earth System Modeling Framework (ESMF). They are a software project that provides a software library to support Earth System modeling. We provide interfaces for ESMF as well as use their regridding capabilities for offline CLM tools.

NCAR National Center for Atmospheric Research (NCAR). This is the research facility that maintains CLM with contributions from other national labs and Universities.

NCEP The National Center for Environmental Prediction (NCEP). In this document this normally refers to the re-analysis atmosphere data produced by NCEP.

PFT Plant Function Type (PFT). A type of vegetation that CLM parameterizes.

PTCLM PoinT CLM (PTCLM) a python script that operates on top of CLM for CLM4.5 to run single point simulations for CLM.

Qian The Qian et. al. analysis of the NCEP forcing data. This can be used to drive CLM with atmosphere forcing from 1948 to 2004. We do NOT expect to be able to update this dataset beyond 2004.

SCRIP Spherical Coordinate Remapping and Interpolation Package (SCRIP). We use it's file format for specifying both grid coordinates as well as mapping between different grids.

VIC Variable Infiltration Capacity (VIC) model for hydrology. This is an option to CLM4.5 in place of the standard CLM4.5 hydrology. The CLM_CONFIG_OPTS option for this is

```
./xmlchange CLM_CONFIG_OPTS="-vichydro on" -append
```

1.2 Setting Up and Running a Case

1.2.1 Choosing a compset

When setting up a new case one of the first choices to make is which “component set” (or compset) to use. The compset refers to which component models are used as well as specific settings for them. We label the different types of compsets with a different letter of the alphabet from “A” (for all data model) to “X” (for all dead model). The compsets of interest when working with CLM are the “I” compsets (which contain CLM with a data atmosphere model and a stub ocean, and stub sea-ice models), “E” and “F” compsets (which contain CLM with the active atmosphere model (CAM), prescribed sea-ice model, and a data ocean model), and “B” compsets which have all active components. Below we go into details on the “I” compsets which emphasize CLM as the only active model, and just mention the two other categories.

When working with CLM you usually want to start with a relevant “I” compset before moving to the more complex cases that involve other active model components. The “I” compsets can exercise CLM in a way that is similar to the coupled modes, but with much lower computational cost and faster turnaround times.

Compsets coupled to data atmosphere and stub ocean/sea-ice (“I” compsets)

Supported CLM Configurations are listed in Table 1-1 for the Scientifically Supported compsets (have been scientifically validated with long simulations) and in Table 1-2 for the Functionally Supported compsets (we’ve only checked that they function).

Scientifically Supported I Compsets:

Short Name	Description	Atm. Forcing	Compset Alias Name	Period
CLM4.5SP	Satellite phenology with new biogeophys	CRUN-CEP	I1850CRU-CLM45	1850
CLM4.5SP	New biogeophys + CENTURY- like vertically resolved soil BGC + CH4 emissions, nitrogen updates	CRUNCEP	I1850CRUCLM45B6IC I20TRCRUCLM420B6Cen-tury	
CLM4.5CN	New biogeophys + CN soil BGC, updates	CRUN-CEP	I1850CRUCLM45B50	
CLM4SP	As in CCSM4/CESM1 release	Qian	I1850	1850
			I	2000
			I20TR	20th Century
CLM4CN	As in CCSM4/CESM1 release	Qian	I1850CN	1850
			ICN	2000
			I20TRCN	20th Century
			IRCP26CN	RCP 2.6 to 2100
			IRCP45CN	RCP 4.5 to 2100
			IRCP60CN	RCP 6.0 to 2100
			IRCP85CN	RCP 8.5 to 2100

Functionally Supported I Compsets:

Short Name	Description	Atm. Forcing	Compset Alias Name	Pe-riod
CLM4.5BGCCROP	ICRU-CLM45BGCCROP	New biogeophys + CENTURY-like vertically resolved soil BGC + CH4 emissions, nitrogen updates with prognostic CROP	CRUNCEP	2000
CLM4.5BGCDV	I1850CRUCLM45B6IDV	New biogeophys + CENTURY-like vertically resolved soil BGC + CH4 emissions, nitrogen updates with DV	CRUNCEP	1850
CLM4.5SP-VIC	ICLM45VIC	Satellite phenology with new biogeophys with VIC hydrology	Qian	2000
CLM4CN-CROP	ICNCROP	As in CCSM4/CESM1 release	Qian	2000
CLM4CN-DV	ICNDV	As in CCSM4/CESM1 release	Qian	1850

Here is the entire list of compsets available. Note that using the “-user_compset” option even more combinations are possible. In the list below we give the alias name and then the long name which describes each component in parenthesis. Alias (Long-name with time-period and each component)

1. I (2000_DATM%QIA_CLM40%SP_SICE_SOCN_RTM_SGLC_SWAV)
2. I1850 (1850_DATM%QIA_CLM40%SP_SICE_SOCN_RTM_SGLC_SWAV)
3. I1850CLM45 (1850_DATM%QIA_CLM45%SP_SICE_SOCN_RTM_SGLC_SWAV)

4. I1850CLM45BGC (1850_DATM%QIA_CLM45%BGC_SICE_SOCN_RTM_SGLC_SWAV)
5. I1850CLM45CN (1850_DATM%QIA_CLM45%CN_SICE_SOCN_RTM_SGLC_SWAV)
6. I1850CLM45CNF (1850_DATM%QIA_CLM45%CN_SICE_SOCN_RTM%FLOOD_SGLC_SWAV)
7. I1850CN (1850_DATM%QIA_CLM40%CN_SICE_SOCN_RTM_SGLC_SWAV)
8. I1850CRU (1850_DATM%CRU_CLM40%SP_SICE_SOCN_RTM_SGLC_SWAV)
9. I1850CRUCLM45 (1850_DATM%CRU_CLM45%SP_SICE_SOCN_RTM_SGLC_SWAV)
10. I1850CRUCLM45BGC (1850_DATM%CRU_CLM45%BGC_SICE_SOCN_RTM_SGLC_SWAV)
11. I1850CRUCLM45BGCDV (1850_DATM%CRU_CLM45%BGCDV_SICE_SOCN_RTM_SGLC_SWAV)
12. I1850CRUCLM45CN (1850_DATM%CRU_CLM45%CN_SICE_SOCN_RTM_SGLC_SWAV)
13. I1850CRUCN (1850_DATM%CRU_CLM40%CN_SICE_SOCN_RTM_SGLC_SWAV)
14. I1850SPINUPCLM45BGC (1850_DATM%S1850_CLM45%BGC_SICE_SOCN_RTM_SGLC_SWAV)
15. I1850SPINUPCN (1850_DATM%S1850_CLM40%CN_SICE_SOCN_RTM_SGLC_SWAV)
16. I1PT (2000_DATM%1PT_CLM40%SP_SICE_SOCN_RTM_SGLC_SWAV)
17. I1PTCLM45 (2000_DATM%1PT_CLM45%SP_SICE_SOCN_RTM_SGLC_SWAV)
18. I20TR (20TR_DATM%QIA_CLM40%SP_SICE_SOCN_RTM_SGLC_SWAV)
19. I20TRCLM45 (20TR_DATM%QIA_CLM45%SP_SICE_SOCN_RTM_SGLC_SWAV)
20. I20TRCLM45CN (20TR_DATM%QIA_CLM45%CN_SICE_SOCN_RTM_SGLC_SWAV)
21. I20TRCN (20TR_DATM%QIA_CLM40%CN_SICE_SOCN_RTM_SGLC_SWAV)
22. I20TRCRU (20TR_DATM%CRU_CLM40%SP_SICE_SOCN_RTM_SGLC_SWAV)
23. I20TRCRUCLM45 (20TR_DATM%CRU_CLM45%SP_SICE_SOCN_RTM_SGLC_SWAV)
24. I20TRCRUCLM45BGC (20TR_DATM%CRU_CLM45%BGC_SICE_SOCN_RTM_SGLC_SWAV)
25. I20TRCRUCLM45CN (20TR_DATM%CRU_CLM45%CN_SICE_SOCN_RTM_SGLC_SWAV)
26. I20TRCRUCN (20TR_DATM%CRU_CLM40%CN_SICE_SOCN_RTM_SGLC_SWAV)
27. I4804 (4804_DATM%QIA_CLM40%SP_SICE_SOCN_RTM_SGLC_SWAV)
28. I4804CLM45 (4804_DATM%QIA_CLM45%SP_SICE_SOCN_RTM_SGLC_SWAV)
29. I4804CLM45CN (4804_DATM%QIA_CLM45%CN_SICE_SOCN_RTM_SGLC_SWAV)
30. I4804CN (4804_DATM%QIA_CLM40%CN_SICE_SOCN_RTM_SGLC_SWAV)
31. ICLM45 (2000_DATM%QIA_CLM45%SP_SICE_SOCN_RTM_SGLC_SWAV)
32. ICLM45BGC (2000_DATM%QIA_CLM45%BGC_SICE_SOCN_RTM_SGLC_SWAV)
33. ICLM45BGCCROP (2000_DATM%QIA_CLM45%BGC-CROP_SICE_SOCN_RTM_SGLC_SWAV)
34. ICLM45BGCDV (2000_DATM%QIA_CLM45%BGCDV_SICE_SOCN_RTM_SGLC_SWAV)
35. ICLM45BGCDVCROP (2000_DATM%QIA_CLM45%BGCDV-CROP_SICE_SOCN_RTM_SGLC_SWAV)
36. ICLM45BGCNOS (2000_DATM%QIA_CLM45%NoVS_SICE_SOCN_RTM_SGLC_SWAV)
37. ICLM45CN (2000_DATM%QIA_CLM45%CN_SICE_SOCN_RTM_SGLC_SWAV)
38. ICLM45CNCROP (2000_DATM%QIA_CLM45%CN-CROP_SICE_SOCN_RTM_SGLC_SWAV)
39. ICLM45CNDV (2000_DATM%QIA_CLM45%CNDV_SICE_SOCN_RTM_SGLC_SWAV)

40. `ICLM45CNTEST` (2003_DATM%QIA_CLM45%CN_SICE_SOCN RTM_SGLC_SWAV_TEST)
41. `ICLM45CRUBGC` (2000_DATM%CRU_CLM45%BGC_SICE_SOCN RTM_SGLC_SWAV)
42. `ICLM45GLCMEC` (2000_DATM%QIA_CLM45%CN_SICE_SOCN RTM_CISM1_SWAV_TEST)
43. `ICLM45SNCRFRC` (2000_DATM%QIA_CLM45%SP-SNCR_SICE_SOCN RTM_SGLC_SWAV)
44. `ICLM45USUMB` (2000_DATM%1PT_CLM45%SP_SICE_SOCN RTM_SGLC_SWAV_CLMUSRDAT%1x1_US-UMB)
45. `ICLM45VIC` (2000_DATM%QIA_CLM45%SP-VIC_SICE_SOCN RTM_SGLC_SWAV)
46. `ICLM45alaskacN` (2000_DATM%QIA_CLM45%CN_SICE_SOCN RTM_SGLC_SWAV_CLMUSRDAT%13x12pt_f19_alas)
47. `ICN` (2000_DATM%QIA_CLM40%CN_SICE_SOCN RTM_SGLC_SWAV)
48. `ICNCROP` (2000_DATM%QIA_CLM40%CN-CROP_SICE_SOCN RTM_SGLC_SWAV)
49. `ICNDV` (2000_DATM%QIA_CLM40%CNDV_SICE_SOCN RTM_SGLC_SWAV)
50. `ICNDVCROP` (2000_DATM%QIA_CLM40%CNDV-CROP_SICE_SOCN RTM_SGLC_SWAV)
51. `ICNTEST` (2003_DATM%QIA_CLM40%CN_SICE_SOCN RTM_SGLC_SWAV_TEST)
52. `ICRU` (2000_DATM%CRU_CLM40%SP_SICE_SOCN RTM_SGLC_SWAV)
53. `ICRUCLM45` (2000_DATM%CRU_CLM45_SICE_SOCN RTM_SGLC_SWAV)
54. `ICRUCLM45BGC` (2000_DATM%CRU_CLM45%BGC_SICE_SOCN RTM_SGLC_SWAV)
55. `ICRUCLM45BGCCROP` (2000_DATM%CRU_CLM45%BGC-CROP_SICE_SOCN RTM_SGLC_SWAV)
56. `ICRUCLM45BGCTEST` (2003_DATM%CRU_CLM45%BGC_SICE_SOCN RTM_SGLC_SWAV_TEST)
57. `ICRUCLM45CN` (2000_DATM%CRU_CLM45%CN_SICE_SOCN RTM_SGLC_SWAV)
58. `ICRUCN` (2000_DATM%CRU_CLM40%CN_SICE_SOCN RTM_SGLC_SWAV)
59. `IG` (2000_DATM%QIA_CLM40%SP_SICE_SOCN RTM_CISM1_SWAV)
60. `IG1850` (1850_DATM%QIA_CLM40%SP_SICE_SOCN RTM_CISM1_SWAV)
61. `IG1850CLM45` (1850_DATM%QIA_CLM45%SP_SICE_SOCN RTM_CISM1_SWAV)
62. `IG1850CLM45CN` (1850_DATM%QIA_CLM45%CN_SICE_SOCN RTM_CISM1_SWAV)
63. `IG1850CN` (1850_DATM%QIA_CLM40%CN_SICE_SOCN RTM_CISM1_SWAV)
64. `IG20TR` (20TR_DATM%QIA_CLM40%SP_SICE_SOCN RTM_CISM1_SWAV)
65. `IG20TRCLM45` (20TR_DATM%QIA_CLM45%SP_SICE_SOCN RTM_CISM1_SWAV)
66. `IG20TRCLM45CN` (20TR_DATM%QIA_CLM45%CN_SICE_SOCN RTM_CISM1_SWAV)
67. `IG20TRCN` (20TR_DATM%QIA_CLM40%CN_SICE_SOCN RTM_CISM1_SWAV)
68. `IG4804` (4804_DATM%QIA_CLM40%SP_SICE_SOCN RTM_CISM1_SWAV)
69. `IG4804CLM45` (4804_DATM%QIA_CLM45%SP_SICE_SOCN RTM_CISM1_SWAV)
70. `IG4804CLM45CN` (4804_DATM%QIA_CLM45%CN_SICE_SOCN RTM_CISM1_SWAV)
71. `IG4804CN` (4804_DATM%QIA_CLM40%CN_SICE_SOCN RTM_CISM1_SWAV)
72. `IGCLM45` (2000_DATM%QIA_CLM45%SP_SICE_SOCN RTM_CISM1_SWAV)
73. `IGCLM45CN` (2000_DATM%QIA_CLM45%CN_SICE_SOCN RTM_CISM1_SWAV)
74. `IGCN` (2000_DATM%QIA_CLM40%CN_SICE_SOCN RTM_CISM1_SWAV)

75. `IGLCMEC (2000_DATM%QIA_CLM40%CN_SICE_SOCN_RTM_CISM1_SWAV_TEST)`
76. `IGRCP26CLM45CN (RCP2_DATM%QIA_CLM45%CN_SICE_SOCN_RTM_CISM1_SWAV)`
77. `IGRCP26CN (RCP2_DATM%QIA_CLM40%CN_SICE_SOCN_RTM_CISM1_SWAV)`
78. `IGRCP45CLM45CN (RCP4_DATM%QIA_CLM45%CN_SICE_SOCN_RTM_CISM1_SWAV)`
79. `IGRCP45CN (RCP4_DATM%QIA_CLM40%CN_SICE_SOCN_RTM_CISM1_SWAV)`
80. `IGRCP60CLM45CN (RCP6_DATM%QIA_CLM45%CN_SICE_SOCN_RTM_CISM1_SWAV)`
81. `IGRCP60CN (RCP6_DATM%QIA_CLM40%CN_SICE_SOCN_RTM_CISM1_SWAV)`
82. `IGRCP85CLM45CN (RCP8_DATM%QIA_CLM45%CN_SICE_SOCN_RTM_CISM1_SWAV)`
83. `IGRCP85CN (RCP8_DATM%QIA_CLM40%CN_SICE_SOCN_RTM_CISM1_SWAV)`
84. `IRCP26CLM45CN (RCP2_DATM%QIA_CLM45%CN_SICE_SOCN_RTM_SGLC_SWAV)`
85. `IRCP26CN (RCP2_DATM%QIA_CLM40%CN_SICE_SOCN_RTM_SGLC_SWAV)`
86. `IRCP45CLM45CN (RCP4_DATM%QIA_CLM45%CN_SICE_SOCN_RTM_SGLC_SWAV)`
87. `IRCP45CN (RCP4_DATM%QIA_CLM40%CN_SICE_SOCN_RTM_SGLC_SWAV)`
88. `IRCP60CLM45CN (RCP6_DATM%QIA_CLM45%CN_SICE_SOCN_RTM_SGLC_SWAV)`
89. `IRCP60CN (RCP6_DATM%QIA_CLM40%CN_SICE_SOCN_RTM_SGLC_SWAV)`
90. `IRCP85CLM45CN (RCP8_DATM%QIA_CLM45%CN_SICE_SOCN_RTM_SGLC_SWAV)`
91. `IRCP85CN (RCP8_DATM%QIA_CLM40%CN_SICE_SOCN_RTM_SGLC_SWAV)`
92. `ISNCRFRC (2000_DATM%QIA_CLM40%SP-SNCR_SICE_SOCN_RTM_SGLC_SWAV)`
93. `ITEST (2003_DATM%QIA_CLM40%SP_SICE_SOCN_RTM_SGLC_SWAV_TEST)`
94. `ITESTCLM45 (2003_DATM%QIA_CLM45%SP_SICE_SOCN_RTM_SGLC_SWAV_TEST)`
95. `IUSUMB (2000_DATM%1PT_CLM40%SP_SICE_SOCN_RTM_SGLC_SWAV_CLMUSRDAT%1x1_US-UMB)`
96. `IalaskaCN (2000_DATM%QIA_CLM40%CN_SICE_SOCN_RTM_SGLC_SWAV_CLMUSRDAT%13x12pt_f19_alaskaUSA)`

Compsets coupled to active atmosphere with data ocean

CAM compsets are compsets that start with “E” or “F” in the name. They are described more fully in the scripts documentation or the CAM documentation. “E” compsets have a slab ocean model while “F” compsets have a data ocean model.

Fully coupled compsets with fully active ocean, sea-ice, and atmosphere

Fully coupled compsets are compsets that start with “B” in the name. They are described more fully in the scripts documentation.

Conclusion to choosing a compset

We’ve introduced the basic type of compsets that use CLM and given some further details for the “standalone CLM” (or “I” compsets). The `config_compsets.xml` lists all of the compsets and gives a full description of each of them. In the next section we look into customizing the setup time options for compsets using CLM.

1.2.2 Customizing CLM's Configuration

The “Creating a Case” section of the CESM1.2.0 Scripts User’s-Guide gives instructions on creating a case. What is of interest here is how to customize your use of CLM for the case that you created.

For CLM when `preview_namelist`, `$CASE.build`, or `$CASE.run` are called there are two steps that take place:

1. The CLM “**configure**” script is called to setup the build-time configuration for CLM (more information on **configure** is given in the Section called [More information on the CLM configure script](#)). The env variables for **configure** are locked after the `$CASE.build` step. So the results of the CLM **configure** are locked after the build has taken place.
2. The CLM “**build-namelist**” script is called to generate the run-time namelist for CLM (more information on **build-namelist** is given below in the Section called [Definition of Namelist items and their default values](#)).

When customizing your case at the `8cesm_setup*` step you are able to modify the process by effecting either one or both of these steps. The CLM “**configure**” and “**build-namelist**” scripts are both available in the “models/lnd/clm/bld” directory in the distribution. Both of these scripts have a “-help” option that is useful to examine to see what types of options you can give either of them.

There are five different types of customization for the configuration that we will discuss: CLM4.5 in CESM1.2.0 build-time options, CLM4.5 in CESM1.2.0 run-time options, User Namelist, other noteworthy CESM1.2.0 configuration items, the CLM **configure** script options, and the CLM **build-namelist** script options.

Information on all of the script, configuration, build and run items is found under `scripts/ccsm_utils/Case.template` in the `config_definition.xml` file.

CLM Script configuration items

Below we list each of the CESM configuration items that are specific to CLM. All of these are available in your: `env_build.xml` and `env_run.xml` files.

```
CLM_CONFIG_OPTS
CLM_BLDNML_OPTS
CLM_NAMELIST_OPTS
CLM_FORCE_COLDSTART
CLM_NML_USE_CASE
CLM_USRDAT_NAME
CLM_CO2_TYPE
```

For the precedence of the different options to **build-namelist** see the section on precedence below.

The first item `CLM_CONFIG_OPTS` has to do with customizing the CLM build-time options for your case, the rest all have to do with generating the namelist.

CLM_CONFIG_OPTS The option `CLM_CONFIG_OPTS` is all about passing command line arguments to the CLM **configure** script. It is important to note that some compsets, may already put a value into the `CLM_CONFIG_OPTS` variable. You can still add more options to your `CLM_CONFIG_OPTS` but make sure you add to what is already there rather than replacing it. Hence, we recommend using the “-append” option to the `xmlchange` script. In the Section called [More information on the CLM configure script](#) below we will go into more details on options that can be customized in the CLM “**configure**” script. It’s also important to note that the `clm.buildnml.csh` script may already invoke certain CLM **configure** options and as such those command line options are NOT going to be available to change at this step (nor would you want to change them). The options to CLM **configure** are given with the “-help” option which is given in the Section called [More information on the CLM configure script](#). .. note:: `CLM_CONFIG_OPTS` is locked after the `$CASE.build` script is run. If you want to change something in `CLM_CONFIG_OPTS` you’ll need to clean the build and rerun `$CASE.build`. The other env variables can be changed at run-time so are never locked.

CLM_NML_USE_CASE CLM_NML_USE_CASE is used to set a particular set of conditions that set multiple namelist items, all centering around a particular usage of the model. To list the valid options do the following:

```
> cd models/lnd/clm/doc  
> ../../bld/build-namelist -use_case list
```

The output of the above command is:

```
CLM build-namelist - use cases: 1850-2100_rcp2.6_glacierMEC_transient 1850-2100_\
˓→rcp2.6_transient \
1850-2100_rcp4.5_glacierMEC_transient 1850-2100_rcp4.5_transient \
1850-2100_rcp6_glacierMEC_transient 1850-2100_rcp6_transient \
1850-2100_rcp8.5_glacierMEC_transient 1850-2100_rcp8.5_transient 1850_control \
1850_glacierMEC_control 2000-2100_rcp8.5_transient 2000_control 2000_glacierMEC_\
˓→control \
20thC_glacierMEC_transient 20thC_transient glacierMEC_pd stdurbpt_pd
Use cases are:....  
  
1850-2100_rcp2.6_glacierMEC_transient = Simulate transient land-use, and aerosol_\
˓→deposition changes \
with historical data from 1850 to 2005 and then with the RCP2.6 scenario from_\
˓→IMAGE  
  
1850-2100_rcp2.6_transient = Simulate transient land-use, and aerosol deposition_\
˓→changes with \
historical data from 1850 to 2005 and then with the RCP2.6 scenario from IMAGE  
  
1850-2100_rcp4.5_glacierMEC_transient = Simulate transient land-use, and aerosol_\
˓→deposition changes \
with historical data from 1850 to 2005 and then with the RCP4.5 scenario from_\
˓→MINICAM  
  
1850-2100_rcp4.5_transient = Simulate transient land-use, and aerosol deposition_\
˓→changes with \
historical data from 1850 to 2005 and then with the RCP4.5 scenario from MINICAM  
  
1850-2100_rcp6_glacierMEC_transient = Simulate transient land-use, and aerosol_\
˓→deposition changes \
with historical data from 1850 to 2005 and then with the RCP6 scenario from AIM  
  
1850-2100_rcp6_transient = Simulate transient land-use, and aerosol deposition_\
˓→changes with \
historical data from 1850 to 2005 and then with the RCP6 scenario from AIM  
  
1850-2100_rcp8.5_glacierMEC_transient = Simulate transient land-use, and aerosol_\
˓→deposition changes \
with historical data from 1850 to 2005 and then with the RCP8.5 scenario from_\
˓→MESSAGE  
  
1850-2100_rcp8.5_transient = Simulate transient land-use, and aerosol deposition_\
˓→changes with \
historical data from 1850 to 2005 and then with the RCP8.5 scenario from MESSAGE  
  
1850_control = Conditions to simulate 1850 land-use  
1850_glacierMEC_control = Running an IG case for 1850 conditions with the ice_\
˓→sheet model glimmer  
2000-2100_rcp8.5_transient = Simulate transient land-use, and aerosol deposition_\
˓→changes with \
```

```

historical data from 2000 to 2005 and then with the RCP8.5 scenario from MESSAGE

2000_control = Conditions to simulate 2000 land-use
2000_glacierMEC_control = Running an IG case for 2000 conditions with the ice_
↳sheet model glimmer
20thC_glacierMEC_transient = Simulate transient land-use, and aerosol deposition_
↳changes from 1850 \
to 2005
20thC_transient = Simulate transient land-use, and aerosol deposition changes_
↳from 1850 to 2005
glacierMEC_pd = Running an IG case with the ice sheet model glimmer
stdurbpt_pd = Standard Urban Point Namelist Settings

.. note:: See `the Section called Precedence of Options <CLM-URL>`_ section for_
↳the precedence of this option relative to the others.

```

CLM_BLDNML_OPTS The option CLM_BLDNML_OPTS is for passing options to the CLM “build-namelist” script. As with the CLM “configure” script the CLM clm.buildnml.csh may already invoke certain options and as such those options will NOT be available to be set here. The best way to see what options can be sent to the “build-namelist” script is to do

```

> cd models/lnd/clm/bld
> ./build-namelist -help

```

Here is the output from the above.

```

./SYNOPSIS
build-namelist [options]

Create the namelist for CLM
OPTIONS
-[no-]chk_res           Also check [do NOT check] to make sure the resolution and
                           land-mask is valid.
-clm_demand "list"      List of variables to require on clm namelist besides the_
                           ↳usuals.
                           "-clm_demand list" to list valid options.
                           (can include a list member "null" which does nothing)
-clm_startfile "file"   CLM restart file to start from.
-clm_start_type "type"  Start type of simulation
                           (default, cold, arb_ic, startup, continue, or branch)
                           (default=do the default type for this configuration)
                           (cold=always start with arbitrary initial conditions)
                           (arb_ic=start with arbitrary initial conditions if
                           initial conditions don't exist)
                           (startup=ensure that initial conditions are being used)
-clm_usr_name    "name" Dataset resolutiondescriptor for personal datasets.
                           Default: not used
                           Example: 1x1pt_boulderCO_c090722 to describe location,
                           number of pts, and date files created
-co2_type "value"       Set CO2 the type of CO2 variation to use.
-co2_ppmv "value"      Set CO2 concentration to use when co2_type is constant_
                           ↳(ppmv).
-config "filepath"     Read the given CLM configuration cache file.
                           Default: "config_cache.xml".
-csmdata "dir"         Root directory of CESM input data.
                           Can also be set by using the CSMDATA environment_
                           ↳variable.
-d "directory"         Directory where output namelist file will be written

```

```

-drydep                               Default: current working directory.
                                         Produce a drydep_inparm namelist that will go into the
                                         "drv flds_in" file for the driver to pass dry-deposition
                                         ↵ to the atm.

                                         Default: -no-drydep
                                         (Note: buildnml.csh copies the file for use by the
                                         ↵ driver)

-glc_grid "grid"                     Glacier model grid and resolution when glacier model,
                                         Only used if glc_nec > 0 for determining fglcmask
                                         Default: gland5UM
                                         (i.e. gland20, gland10 etcetera)

-glc_nec <name>                   Glacier number of elevation classes [0 | 3 | 5 | 10 | 36]
                                         (default is 0) (standard option with land-ice model is
                                         ↵ 10)

-glcsmb <value>                  Only used if glc_nec > 0
                                         If .true., pass surface mass balance info to GLC
                                         If .false., pass positive-degree-day info to GLC
                                         Default: true
                                         Print usage to STDOUT.

-ignore_ic_date                     Ignore the date on the initial condition files
                                         when determining what input initial condition file to
                                         ↵ use.

-ignore_ic_year                    Ignore just the year part of the date on the initial
                                         ↵ condition files
                                         when determining what input initial condition file to
                                         ↵ use.

-infile "filepath"                 Specify a file (or list of files) containing namelists to
                                         read values from.

                                         If used with a CLM build with multiple ensembles (ninst_
                                         ↵ lnd>1)
                                         and the filename entered is a directory to files of the
                                         form filepath/filepath and filepath/filepath_$n where $n
                                         is the ensemble member number. the "filepath/filepath"
                                         input namelist file is the master input namelist file
                                         that is applied to ALL ensemble members.

                                         (by default for CESM this is setup for files of the
                                         form $CASEDIR/user_nl_clm/user_nl_clm_????)
                                         Writes out a list containing pathnames for required
                                         ↵ input datasets in
                                         file specified.

-irrig "value"                      If .true. turn irrigation on with namelist logical
                                         ↵ irrigate (for CLM4.5 physics)
                                         (requires crop to be on in the clm configuration)
                                         Seek surface datasets with irrigation turned on. (for
                                         ↵ CLM4.0 physics)
                                         Default: .false.
                                         Number of CLM coupling time-steps in a day.
                                         Land fraction file (the input domain file)
                                         Type of land-mask (default, navy, gx3v5, gx1v5 etc.)
                                         "-mask list" to list valid land masks.
                                         Specify namelist settings directly on the commandline by
                                         ↵ supplying
                                         a string containing FORTRAN namelist syntax, e.g.,
                                         -namelist "&clm_inparm dt=1800 /"
                                         DO NOT PRODUCE a megan_emis_nl namelist that will go
                                         ↵ into the

```

```

  ↵atm.
  ↵Nature)
  ↵driver)
-[no-]note
-rcp "value"
-res "resolution"
  ↵grids;
  ↵size
-s
-sim_year "year"
-bgc_spinup "on|off"
  ↵configure.
  ↵automatically
  ↵spinup
-test
  ↵filesystem.
-verbose [or -v]
-use_case "case"
-version
  ↵distribution.

"drv flds_in" file for the driver to pass VOCs to the MEGAN (Model of Emissions of Gases and Aerosols from (Note: buildnml.csh copies the file for use by the Add note to output namelist [do NOT add note] about the arguments to build-namelist. Representative concentration pathway (rcp) to use for future scenarios. "-rcp list" to list valid rcp settings. Specify horizontal grid. Use nlatxlon for spectral dlatxlon for fv grids (dlat and dlon are the grid cell in degrees for latitude and longitude respectively) "-res list" to list valid resolutions. Turns on silent mode - only fatal messages issued. Year to simulate for input datasets (i.e. 1850, 2000, 1850-2000, 1850-2100) "-sim_year list" to list valid simulation years CLM 4.5 Only. For CLM 4.0, spinup is controlled from Turn on given spinup mode for BGC setting of CN on : Turn on Accelerated Decomposition (spinup_state = 1) off : run in normal mode (spinup_state = 0) Default is off. Spinup is now a two step procedure. First, run the model with spinup = "on". Then run the model for a while with spinup = "off". The exit spinup step happens on the first timestep when using a restart file from mode. The spinup state is saved to the restart file. If the values match between the model and the restart file it proceeds as directed. If the restart file is in spinup mode and the model is in normal mode, then it performs the exit spinup step and proceeds in normal mode after that. If the restart file has normal mode and the model is in spinup, then it enters spinup. This is useful if you a parameter and want to rapidly re-equilibrate without a cold start. Enable checking that input datasets exist on local Turn on verbose echoing of informational messages. Specify a use case which will provide default values. "-use_case list" to list valid use-cases. Echo the SVN tag name used to check out this CLM

```

```
Note: The precedence for setting the values of namelist variables is (highest to lowest):
0. namelist values set by specific command-line options, like, -d, -sim_year
   (i.e. CLM_BLDNML_OPTS env_run variable)
1. values set on the command-line using the -namelist option,
   (i.e. CLM_NAMELIST_OPTS env_run variable)
2. values read from the file(s) specified by -infile,
   (i.e. user_nl_clm files)
3. datasets from the -clm_usr_name option,
   (i.e. CLM_USRDATA_NAME env_run variable)
4. values set from a use-case scenario, e.g., -use_case
   (i.e. CLM_NML_USE_CASE env_run variable)
5. values from the namelist defaults file.
```

The **clm.buildnml.csh** script already sets the resolution and mask as well as the CLM **configure** file, and defines an input namelist and namelist input file, and the output namelist directory, and sets the start-type (from `RUN_TYPE`), namelist options (from `CLM_NAMELIST_OPTS`), `co2_ppmv` (from `CCSM_CO2_PPMV`, `co2_type` (from `CLM_CO2_TYPE`), `lnd_frac` (from `LND_DOMAIN_PATH` and `LND_DOMAIN_FILE`), `l_ncpl` (from `LND_NCPL`, `glc_grid`, `glc_smb`, `glc_nec` (from `GLC_GRID`, `GLC_SMB`, and `GLC_NECK`), and “`clm_usr_name`” is set (to `CLM_USRDATA_NAME > ``` when the grid is set to ```CLM_USRDATA_NAME`. Hence only the following different options can be set:

1. `-bgc_spinup`
2. `-chk_res`
3. `-clm_demand`
4. `-drydep`
5. `-ignore_ic_date`
6. `-ignore_ic_year`
7. `-irrig`
8. `-no-megan`
9. `-note`
10. `-rcp`
11. `-sim_year`

“`-bgc_spinup`” is an option only available for CLM4.5 for any configuration when CN is turned on (so either CLMCN or CLMBGC). It can be set to “on” or “off”. If “on” the model will go into Accelerated Decomposition mode, while for “off” (the default) it will have standard decomposition rates. If you are starting up from initial condition files the model will check what mode the initial condition file is in and do the appropriate action on the first time-step to change the Carbon pools to the appropriate spinup setting. See the Section called Spinning up the CLM4.5 biogeochemistry (CLMBGC spinup) in Chapter 4 for an example using this option.

“`-chk_res`” ensures that the resolution chosen is supported by CLM. If the resolution is NOT supported it will cause the CLM **build-namelist** to abort when run. So when either `preview_namelist`, `$CASE.build` or `$CASE.run` is executed it will abort early. Since, the CESM scripts only support certain resolutions anyway, in general this option is NOT needed in the context of running CESM cases.

“-clm_demand” asks the **build-namelist** step to require that the list of variables entered be set. Typically, this is used to require that optional filenames be used and ensure they are set before continuing. For example, you may want to require that fpftdyn be set to get dynamically changing vegetation types. To do this you would do the following.

```
> ./xmlchange CLM_BLDNML_OPTS="-clm_demand fpftdyn"``
```

To see a list of valid variables that you could set do this:

```
> cd models/lnd/clm/doc
> ../bld/build-namelist -clm_demand list
```

Note: Using a 20th-Century transient compset or the 20thC_transient use-case using `CLM_NML_USE_CASE` would set this as well, but would also use dynamic nitrogen and aerosol deposition files, so using `-clm_demand` would be a way to get *just* dynamic vegetation types and NOT the other files as well.

“-drydep” adds the dry-deposition namelist to the driver. This is a driver namelist, but adding the option here has CLM **build-namelist** create the `drv_flds_in` file that the driver will copy over and use. Invoking this option does have an impact on performance even for I compsets and will slow the model down. It’s also only useful when running with an active atmosphere model that makes use of this information.

“-ignore_ic_date” ignores the Initial Conditions (IC) date completely for finding initial condition files to startup from. Without this option or the “-ignore_ic_year” option below, the date of the file comes into play.

“-ignore_ic_year” ignores the Initial Conditions (IC) year for finding initial condition files to startup from. The date is used, but the year is ignored. Without this option or the “-ignore_ic_date” option below, the date and year of the file comes into play.

When “-irrig” is used **build-namelist** will try to find surface datasets that have the irrigation model enabled.

“no-megan” means do NOT add the MEGAN model Biogenic Volatile Organic Compounds (BVOC) namelist to the driver. This namelist is created by default, so normally this WILL be done. This is a driver namelist, so unless “no-megan” is specified the CLM **build-namelist** will create the `drv_flds_in` file that the driver will copy over and use (if you are running with CAM and CAM produces this file as well, it’s file will have precedence).

“-note” adds a note to the bottom of the namelist file, that gives the details of how **build-namelist** was called, giving the specific command-line options given to it.

“-rcp” is used to set the representative concentration pathway for the future scenarios you want the data-sets to simulate conditions for, in the input datasets. To list the valid options do the following:

```
> cd models/lnd/clm/doc
> ../bld/build-namelist -rcp list
```

“-sim_year” is used to set the simulation year you want the data-sets to simulate conditions for in the input datasets. The simulation “year” can also be a range of years in order to do simulations with changes in the dataset values as the simulation progresses. To list the valid options do the following:

```
> cd models/lnd/clm/doc
> ../bld/build-namelist -sim_year list
```

`CLM_NAMELIST_OPTS` passes namelist items into one of the CLM namelists.

Note: For character namelist items you need to use ‘‘’ as quotes for strings so that the scripts don’t get confused with other quotes they use.

Example, you want to set `hist_dov2xy` to `.false.` so that you get vector output to your history files. To do so edit `env_run.xml` and add a setting for `hist_dov2xy`. So do the following:

```
> ./xmlchange CLM_NAMELIST_OPTS="hist_dov2xy=.false."
```

Example, you want to set `hist_fincl1` to add the variable ‘HK’ to your history files. To do so edit `env_run.xml` and add a setting for `hist_fincl1`. So do the following:

```
> ./xmlchange CLM_NAMELIST_OPTS="hist_fincl1='HK'"
```

For a list of the history fields available see [CLM History Fields](#).

CLM_FORCE_COLDSTART when set to on, *requires* that your simulation do a cold start from arbitrary initial conditions. If this is NOT set, it will use an initial condition file if it can find an appropriate one, and otherwise do a cold start. **CLM_FORCE_COLDSTART** is a good way to ensure that you are doing a cold start if that is what you want to do.

CLM_USRDAT_NAME Provides a way to enter your own datasets into the namelist. The files you create must be named with specific naming conventions outlined in: [the Section called Creating your own single-point/regional surface datasets in Chapter 5](#). To see what the expected names of the files are, use the **queryDefaultNamelist.pl** to see what the names will need to be. For example if your `CLM_USRDAT_NAME` will be “`1x1_boulderCO`”, with a “navy” land-mask, constant simulation year range, for 1850, the following will list what your filenames should be:

```
> cd models/lnd/clm/bld
> queryDefaultNamelist.pl -usrname "1x1_boulderCO" -options mask=navy,sim_
  ↵year=1850,sim_year_range="constant" -csmdata $CSMDATA
```

An example of using ```CLM_USRDAT_NAME``` for a simulation is given in `Example 5-4`_.

CLM_CO2_TYPE sets the type of input CO2 for either “constant”, “diagnostic” or prognostic”. If “constant” the value from `CCSM_CO2_PPMV` will be used. If “diagnostic” or “prognostic” the values MUST be sent from the atmosphere model. For more information on how to send CO2 from the data atmosphere model see [the Section called Running stand-alone CLM with transient historical CO2 concentration in Chapter 4](#).

User Namelist

`CLM_NAMELIST_OPTS` as described above allows you to set any extra namelist items you would like to appear in your namelist. However, it only allows you a single line to enter namelist items, and strings must be quoted with `'`; which is a bit awkward. If you have a long list of namelist items you want to set (such as a long list of history fields) a convenient way to do it is to add to the `user_nl_clm` that is created after the `cesm_setup` command runs. The file needs to be in valid FORTRAN namelist format (with the exception that the namelist name &namelist and the end of namelist marker “/” are excluded). The `preview_namelist` or `$CASE.run` step will abort if there are syntax errors. All the variable names must be valid and the values must be valid for the datatype and any restrictions for valid values for that variable. Here’s an example `user_nl_clm` namelist that sets a bunch of history file related items, to create output history files monthly, daily, every six and 1 hours.

Example: `user_nl_clm` namelist file

```
!-----
! Users should add all user specific namelist changes below in the form of
! namelist_var = new_namelist_value
!
```

```

! Include namelist variables for drv flds_in ONLY if -megan and/or -drydep options
! are set in the CLM_NAMELIST_OPTS env variable.
!
! EXCEPTIONS:
! Set co2_ppmv      with CCSM_CO2_PPMV          option
! Set dtime         with L_NCPL                option
! Set fatmlndfrc   with LND_DOMAIN_PATH/LND_DOMAIN_FILE    options
! Set finidat      with RUN_REFCASE/RUN_REFDATE/RUN_REFTOD options for hybrid_
! or branch cases
!                               (includes $inst_string for multi-ensemble cases)
! Set glc_grid       with GLC_GRID              option
! Set glc_smb        with GLC_SMB               option
! Set maxpatch_glcme with GLC_NECK             option
!-----
hist_fincl2 = 'TG', 'TBOT', 'FIRE', 'FIRA', 'FLDS', 'FSDS',
              'FSR', 'FSA', 'FGEV', 'FSH', 'FGR', 'TSOI',
              'ERRSOI', 'BUILDHEAT', 'SABV', 'SABG',
              'FSDSVD', 'FSDSND', 'FSDSVI', 'FSDSNI',
              'FSRVD', 'FSRND', 'FSRVI', 'FSRNI',
              'TSA', 'FCTR', 'FCEV', 'QBOT', 'RH2M', 'H2OSOI',
              'H2OSNO', 'SOILLIQ', 'SOILICE',
              'TSA_U', 'TSA_R',
              'TREFMNAV_U', 'TREFMNAV_R',
              'TREFMXAV_U', 'TREFMXAV_R',
              'TG_U', 'TG_R',
              'RH2M_U', 'RH2M_R',
              'QRUNOFF_U', 'QRUNOFF_R',
              'SoilAlpha_U',
              'Qanth', 'SWup', 'LWup', 'URBAN_AC', 'URBAN_HEAT'
hist_fincl3 = 'TG:I', 'FSA:I', 'SWup:I', 'URBAN_AC:I', 'URBAN_HEAT:I',
              'TG_U:I', 'TG_R:I',
hist_fincl4 = 'TG', 'FSA', 'SWup', 'URBAN_AC', 'URBAN_HEAT'
hist_mfilt = 1, 30, 28, 24
hist_nhtfrq = 0, -24, -6, -1

```

Note: The comments at the top are some guidance given in the default user_nl_clm and just give some guidance on how to set variables and use the file.

Note: See [the Section called Precedence of Options](#) section for the precedence of this option relative to the others.

Note: You do NOT need to specify the namelist group that the variables are in because the CLM **build-namelist** knows the namelist that specific variable names belong to, and it puts them there.

Obviously, all of this would be difficult to put in the CLM_NAMELIST_OPTS variable, especially having to put ' around all the character strings. For more information on the namelist variables being set here and what they mean, see the section on CLM namelists below, as well as the namelist definition that gives details on each variable.

Precedence of Options

Note: The precedence for setting the values of namelist variables with the different env_build.xml, env_run.xml options is (highest to lowest):

1. Namelist values set by specific command-line options, like, -d, -sim_year (i.e. CLM_BLDNML_OPTS env_build.xml variable)
2. Values set on the command-line using the -namelist option, (i.e. CLM_NAMELIST_OPTS env_run.xml variable)

3. Values read from the file specified by `-infile`, (i.e. `user_nl_clm` file)
4. Datasets from the `-clm_usr_name` option, (i.e. `CLM_USRDAT_NAME` `env_run.xml` variable)
5. Values set from a use-case scenario, e.g., `-use_case` (i.e. `CLM_NML_USE_CASE` `env_run.xml` variable)
6. Values from the namelist defaults file.

Thus a setting in `CLM_BLDNML_OPTS` will override a setting for the same thing given in a use case with `CLM_NML_USE_CASE`. Likewise, a setting in `CLM_NAMELIST_OPTS` will override a setting in `user_nl_clm`.

Setting Your Initial Conditions File

Especially with CLMBGC and CLMCN starting from initial conditions is very important. Even with CLMSP it takes many simulation years to get the model fully spunup. There are a couple different ways to provide an initial condition file.

- the Section called Doing a hybrid simulation to provide initial conditions
- the Section called Doing a branch simulation to provide initial conditions
- the Section called Providing a finidat file in your `user_nl_clm` file
- the Section called Adding a finidat file to the XML database

Note: Your initial condition file MUST agree with the surface dataset you are using to run the simulation. If the two files do NOT agree you will get a run-time about a mis-match in PFT weights, or in the number of PFT's or columns. To get around this you'll need to use the Section called Using interpinic to interpolate initial conditions to different resolutions in Chapter 2 to interpolate your initial condition dataset.

Doing a hybrid simulation to provide initial conditions

The first option is to setup a hybrid simulation and give a `RUN_REFCASE` and `RUN_REFDATE` to specify the reference case simulation name to use. When you setup most cases, at the standard resolutions of “f09” or “f19” it will already do this for you. For example, if you run an “I2000CN” compset at “f09_g16” resolution the following settings will already be done for you.

```
./xmlchange RUN_TYPE=hybrid,RUN_REFCASE=I2000CN_f09_g16_c100503,  
RUN_REFDATE=0001-01-01,GET_REFCASE=TRUE
```

Setting the `GET_REFCASE` option to `TRUE` means it will copy the files from the: `$DIN_LOC_ROOT/ccsm4_init/I2000CN_f09_g16_c100503/0001-01-01` directory. Note, that the `RUN_REFCASE` and `RUN_REFDATE` variables are expanded to get the directory name above. If you do NOT set `GET_REFCASE` to `TRUE` then you will need to have placed the file in your run directory yourself. In either case, the file is expected to be named: `$RUN_REFCASE.clm2.r.$RUN_REFDATE-00000.nc` with the variables expanded of course.

Doing a branch simulation to provide initial conditions

The setup for running a branch simulation is essentially the same as for a hybrid. With the exception of setting `RUN_TYPE` to branch rather than hybrid. A branch simulation runs the case essentially as restarting from its place before to exactly reproduce it (but possibly output more or different fields on the history files). While a hybrid simulation allows you to change the configuration or run-time options, as well as use a different code base than the original case that may have fewer fields on it than a full restart file. The `GET_REFCASE` option works similarly for a branch case as for a hybrid.

Providing a finidat file in your user_nl_clm file

Setting up a branch or hybrid simulation requires the initial condition file to follow a standard naming convention, and a standard input directory if you use the GET_REFCASE option. If you want to name your file willy nilly and place it anywhere, you can set it in your `user_nl_clm` file as in this example.

```
finidat      = '/glade/home/$USER/myinitdata/clmi_I1850CN_f09_g16_0182-01-01.cl20329.nc'
```

Note, if you provide an initial condition file – you can NOT set `CLM_FORCE_COLDSTART` to TRUE.

Adding a finidat file to the XML database

Like other datasets, if you want to use a given initial condition file to be used for all (or most of) your cases you'll want to put it in the XML database so it will be used by default. The initial condition files, are resolution dependent, and dependent on the number of PFT's and other variables such as GLC_NECK or if irrigation is on or off. See Chapter 3 for more information on this.

Other noteworthy configuration items

For running “I” cases there are several other noteworthy configuration items that you may want to work with. Most of these involve settings for the DATM, but one `CCSM_CO2_PPMV` applies to all models. If you are running an B, E, or F case that doesn't use the DATM obviously the `DATM_*` settings will not be used. All of the settings below are in your `env_build.xml` and `env_run.xml` files

```
CCSM_CO2_PPMV
CCSM_VOC
DATM_MODE
DATM_PRESAERO
DATM_CLMNCEP_YR_ALIGN
DATM_CLMNCEP_YR_START
DATM_CLMNCEP_YR_END
DATM_CPL_CASE
DATM_CPL_YR_ALIGN
DATM_CPL_YR_START
DATM_CPL_YR_END
```

CCSM_CO2_PPMV Sets the mixing ratio of CO₂ in parts per million by volume for ALL CESM components to use.

Note that most compsets already set this value to something reasonable. Also note that some compsets may tell the atmosphere model to override this value with either historic or ramped values. If the `CCSM_BGC` variable is set to something other than “none” the atmosphere model will determine CO₂, and CLM will listen and use what the atmosphere sends it. On the CLM side the namelist item `co2_type` tells CLM to use the value sent from the atmosphere rather than a value set on its own namelist.

CCSM_VOC Enables passing of the Volatile Organic Compounds (VOC) from CLM to the atmospheric model. This of course is only important if the atmosphere model is a fully active model that can use these fields in its chemistry calculations.

DATM_MODE Sets the mode that the DATM model should run in this determines how data is handled as well as what the source of the data will be. Many of the modes are setup specifically to be used for ocean and/or sea-ice modeling. The modes that are designed for use by CLM are:

```
CLMCRUNCEP
CLM_QIAN
```

```
CLM1PT>
CPLHIST3HrWx
```

CLMCRUNCEP The standard mode for CLM4.5 of using global atmospheric data that was developed by CRU using NCEP data from 1901 to 2010. See the Section called [CLMCRUNCEP mode](#) and it's DATM settings for more information on the DATM settings for CLMCRUNCEP mode.

CLM_QIAN The standard mode for CLM4.0 of using global atmospheric data that was developed by Qian et. al. for CLM using NCEP data from 1948 to 2004. See the Section called [CLM_QIAN mode](#) and it's DATM settings for more information on the DATM settings for CLM_QIAN mode. CLM1PT is for the special cases where we have single-point tower data for particular sites. Right now we only have data for three urban locations: MexicoCity Mexico, Vancouver Canada, and the urban-c alpha site. And we have data for the US-UMB AmeriFlux tower site for University of Michigan Biological Station. See the Section called [CLM1PT mode](#) and it's DATM settings for more information on the DATM settings for CLM1PT mode. CPLHIST3HrWx is for running with atmospheric forcing from a previous CESM simulation. See the Section called [CPLHIST3HrWx mode](#) and it's DATM settings for more information on the DATM settings for CPLHIST3HrWx mode.

DATM_PRESAERO sets the prescribed aerosol mode for the data atmosphere model. The list of valid options include:

```
clim_1850 = constant year 1850 conditions
clim_2000 = constant year 2000 conditions
trans_1850-2000 = transient 1850 to year 2000 conditions
rcp2 . 6 = transient conditions for the rcp=2.6 W/m2 future scenario
rcp4 . 5 = transient conditions for the rcp=4.5 W/m2 future scenario
rcp6 . 0 = transient conditions for the rcp=6.0 W/m2 future scenario
rcp8 . 5 = transient conditions for the rcp=8.5 W/m2 future scenario
pt1_pt1 = read in single-point or regional datasets
```

DATM_CLMNCEP_YR_START DATM_CLMNCEP_YR_START sets the beginning year to cycle the atmospheric data over for CLM_QIAN or CLMCRUNCEP modes.

DATM_CLMNCEP_YR_END DATM_CLMNCEP_YR_END sets the ending year to cycle the atmospheric data over for CLM_QIAN or CLMCRUNCEP modes.

DATM_CLMNCEP_YR_ALIGN DATM_CLMNCEP_YR_START and DATM_CLMNCEP_YR_END determine the range of years to cycle the atmospheric data over, and DATM_CLMNCEP_YR_ALIGN determines which year in that range of years the simulation will start with.

DATM_CPL_CASE DATM_CPL_CASE sets the casename to use for the CPLHIST3HrWx mode.

DATM_CPL_YR_START DATM_CPL_YR_START sets the beginning year to cycle the atmospheric data over for the CPLHIST3HrWx mode.

DATM_CPL_YR_END DATM_CPL_YR_END sets the ending year to cycle the atmospheric data over for the CPLHIST3HrWx mode.

DATM_CPL_YR_ALIGN DATM_CPL_YR_START and DATM_CPL_YR_END determine the range of years to cycle the atmospheric data over, and DATM_CPL_YR_ALIGN determines which year in that range of years the simulation will start with.

Downloading DATM Forcing Data

In Chapter One of the [CESM User's Guide](#) there is a section on “Downloading input data”. The normal process of setting up cases will use the “scripts/ccsm_utils/Tools/check_input_data” script to retrieve data from the CESM

subversion inputdata repository. This is true for the standard *CLM_QIAN* forcing as well.

The *CLMCRUNCEP* data is uploaded into the subversion inputdata repository as well – but as it is 1.1 Terabytes of data downloading it is problematic (*IT WILL TAKE SEVERAL DAYS TO DOWNLOAD THE ENTIRE DATASET USING SUBVERSION*). Because of its size you may also need to download it onto a separate disk space. We have done that on yellowstone for example where it resides in `$ENV{CESMROOT}/lmwg` while the rest of the input data resides in `$ENV{CESMDATAROOT}/inputdata`. The data is also already available on: janus, franklin, and hopper. If you download the data, we recommend that you break your download into several chunks, by setting up a case and setting the year range for `DATM_CPL_YR_START` and `DATM_CPL_YR_END` in say 20 year sections over 1901 to 2010, and then use `check_input_data` to export the data.

The `CPLHIST3HrWx` DATM forcing data is unique – because it is large compared to the rest of the input data, and we only have a disk copy on yellowstone. The DATM assumes the path for the previous NCAR machine yellowstone of `/glade/p/cesm/shared_outputdata/cases/ccsm4/$DATM_CPLHIST_CASE` for the data. So you will need to change this path in order to run on any other machine. You can download the data itself from NCAR HPSS from `/CCSM/csm/$DATM_CPLHIST_CASE`.

Customizing via the build script files

The final thing that the user may wish to do before `cesm_setup` is run is to edit the build script files which determine the configuration and namelist. The variables in `env_build.xml` or `env_run.xml` typically mean you will NOT have to edit build script files. But, there are rare instances where it is useful to do so. The build script files are copied to your case directory and are available under `Buildconf`. The list of build script files you might wish to edit are:

`clm.buildexe.csh` `clm.buildnml.csh` `datm.buildexe.csh` `datm.buildnml.csh`

More information on the CLM configure script

The CLM **configure** script defines the details of a clm configuration and summarizes it into a `config_cache.xml` file. The `config_cache.xml` will be placed in your case directory under `Buildconf/clmconf`. The `config_definition.xml` in `models/lnd/clm/bld/config_files` gives a definition of each CLM configuration item, it is viewable in a web-browser. Many of these items are things that you would NOT change, but looking through the list gives you the valid options, and a good description of each. Below we repeat the `config_definition.xml` file contents:

Help on CLM configure

Coupling this with looking at the options to CLM **configure** with “-help” as below will enable you to understand how to set the different options.

```
> cd models/lnd/clm/bld
> configure -help
```

The output to the above command is as follows:

```
SYNOPSIS
  configure [options]

  Configure CLM in preparation to be built.

OPTIONS
  User supplied values are denoted in angle brackets (<>). Any value that contains
  white-space must be quoted. Long option names may be supplied with either single
  or double leading dashes. A consequence of this is that single letter options may
  NOT be bundled.
```

```

-bgc <name>          Build CLM with BGC package [ none | cn | cndv ]
                      (default is none).
-cache <file>        Name of output cache file (default: config_cache.xml).
-cachedir <file>     Name of directory where output cache file is written
                      (default: CLM build directory).
-clm4me <name>       Turn Methane model: [on | off]
                      Requires bgc=cn/cndv (Carbon Nitrogen model)
                      (ONLY valid for CLM4.5!)
-clm_root <dir>      Root directory of clm source code
                      (default: directory above location of this script)
-cppdefs <string>    A string of user specified CPP defines. Appended to
                      Makefile defaults. e.g. -cppdefs '-DVAR1 -DVAR2'
-vichydro <name>     Turn VIC hydrologic parameterizations : [on | off] (default
-is off)              ↵
-crop <name>         Toggle for prognostic crop model. [on | off] (default is off)
                      (can ONLY be turned on when BGC type is CN or CNDV)
-comp_intf <name>   Component interface to use (ESMF or MCT) (default MCT)
-defaults <file>    Specify full path to a configuration file which will be used
                      to supply defaults instead of the defaults in bld/config_
files.                ↵
only.                 ↵
will.                ↵
-exlaklayers <name> Turn on extra lake layers (25 layers instead of 10) [on | off]
-off]                 ↵
(ONLY valid for CLM4.5!)
-help [or -h]         Print usage to STDOUT.
-nofire               Turn off wildfires for BGC setting of CN
                      (default includes fire for CN)
-noio                 Turn history output completely off (typically for testing).
-phys <name>         Value of clm4_0 or clm4_5 (default is clm4_0)
-silent [or -s]       Turns on silent mode - only fatal messages issued.
-sitespf_pt <name>  Setup for the given site specific single-point resolution.
-snifar_frc <name>  Turn on SNICAR radiative forcing calculation. [on | off]
                      (default is off)
-spinup <name>      CLM 4.0 Only. For CLM 4.5, spinup is controlled from build-
namelist.             ↵
Turn on given spinup mode for BGC setting of CN ↵
↳ (level)
  ↳ (2)               AD           Turn on Accelerated Decomposition from ↵
                        bare-soil
                        exit         Jump directly from AD spinup to normal mode ↵
  ↳ (1)               normal        Normal decomposition ("final spinup mode") ↵
  ↳ (0)               (default)
                        The recommended sequence is 2-1-0
-usr_src <dir1>[,<dir2>[,<dir3>[...]]] Directories containing user source code.
-verbose [or -v]      Turn on verbose echoing of settings made by configure.
-version              Echo the SVN tag name used to check out this CLM
distribution.          ↵
-vsoilc_centbgc <name> Turn on vertical soil Carbon profile, CENTURY model
-decomposition, \

```

```

split Nitrification/de-Nitrification into two mineral
pools for NO3 and NH4 (requires clm4me Methane model), and
eliminate inconsistent duplicate soil hydraulic
parameters used in soil biogeochem.
(requires either CN or CNDV)
(ONLY valid for CLM4.5!)
[on,off or colon delimited list of no options] (default off)
    no-vert      Turn vertical soil Carbon profile off
    no-cent     Turn CENTURY off
    no-nitrif   Turn the Nitrification/denitrification off
[no-vert,no-cent,no-nitrif,no-vert:no-cent]

```

We've given details on how to use the options in env_build.xml and env_run.xml to interact with the CLM "configure" and "build-namelist" scripts, as well as giving a good understanding of how these scripts work and the options to them. In the next section we give further details on the CLM namelist. You could customize the namelist for these options after "cesm_setup" is run.

1.2.3 Customizing CLM's namelist

Once a case is **cesm_setup**, we can then customize the case further, by editing the run-time namelist for CLM. First let's list the definition of each namelist item and their valid values, and then we'll list the default values for them. Next for some of the most used or tricky namelist items we'll give examples of their use, and give you example namelists that highlight these features.

In the following, various examples of namelists are provided that feature the use of different namelist options to customize a case for particular uses. Most the examples revolve around how to customize the output history fields. This should give you a good basis for setting up your own CLM namelist.

Definition of Namelist items and their default values

Here we point to you where you can find the definition of each namelist item and separately the default values for them. The default values may change depending on the resolution, land-mask, simulation-year and other attributes. Both of these files are viewable in your web browser, and then expand each in turn.

1. Definition of Namelists Relevant for CLM4.5
2. Default values of each CLM4.0 Namelist Item
3. Default values of each CLM4.5 Namelist Item

List of fields that can be added to your output history files by namelist

One set of the namelist items allows you to add fields to the output history files: hist_fincl1, hist_fincl2, hist_fincl3, hist_fincl4, hist_fincl5, and hist_fincl6. The following links for [CLM4.0 History Fields](#) and [CLM4.5 History Fields](#) documents all of the history fields available and gives the long-name and units for each. The table below lists all the CLM4.5 history fields.

Definition of CLM history variables

Included in the table are the following pieces of information:

- Variable name.

- Long name description.
- units

Table 1-3. CLM History Fields

Table goes here.

Examples of using different namelist features

Below we will give examples of user namelists that activate different commonly used namelist features. We will discuss the namelist features in different examples and then show a user namelist that includes an example of the use of these features. First we will show the default namelist that doesn't activate any user options.

The default namelist

Here we give the default namelist as it would be created for an “I1850CRUCLM45BGC” compset at 0.9x1.25 resolution with a gx1v6 land-mask on yellowstone. To edit the namelist you would edit the `user_nl_clm` user namelist with just the items you want to change. For simplicity we will just show the CLM namelist and NOT the entire file. In the sections below, for simplicity we will just show the user namelist (`user_nl_clm`) that will add (or modify existing) namelist items to the namelist.

Example 1-2. Default CLM Namelist

```
&clm_inparm
albice = 0.60,0.40
co2_ppmv = 284.7
co2_type = 'constant'
create_crop_landunit = .false.
dtime = 1800
fatmlndfrc = '/glade/p/cesm/cseg/inputdata/share/domains/domain.lnd.fv0.9x1.25_gx1v6.
˓→090309.nc'
finidat = '$DIN_LOC_ROOT/lnd/clm2/initdata_map/clmi.I1850CRUCLM45BGC.0241-01-01.0.
˓→9x1.25_gx1v6_simyr1850_c130531.nc'
fpftcon = '/glade/p/cesm/cseg/inputdata/lnd/clm2/pftdata/pft-physiology.c130503.nc'
fsnowaging = '/glade/p/cesm/cseg/inputdata/lnd/clm2/snicalldata/snical_drift_bst_fit_
˓→60_c070416.nc'
fsnowoptics = '/glade/p/cesm/cseg/inputdata/lnd/clm2/snicalldata/snical_optics_5bnd_
˓→c090915.nc'
fsurdat = '/glade/p/cesm/cseg/inputdata/lnd/clm2/surfdata_map/surfdata_0.9x1.25_
˓→simyr1850_c130415.nc'
maxpatch_glcme = 0
more_vertlayers = .false.
nsegspc = 20
spinup_state = 0
urban_hac = 'ON'
urban_traffic = .false.
/
&ndepdyn_nml
ndepmapalgo = 'bilinear'
stream_fldfilename_ndep = '/glade/p/cesm/cseg/inputdata/lnd/clm2/ndepdata/fndep_clm_
˓→hist_simyr1849-2006_1.9x2.5_c100428.nc'
stream_year_first_ndep = 1850
```

```

stream_year_last_ndep = 1850
/
&popd_streams
popdensmapalgo = 'bilinear'
stream_fldfilename_popdens = '$DIN_LOC_ROOT/lnd/clm2/firedata/clmforc.Li_2012_hdm_0.
↪5x0.5_AVHRR_simyr1850-2010_c130401.nc'
stream_year_first_popdens = 1850
stream_year_last_popdens = 1850
/
&light_streams
lightngmapalgo = 'bilinear'
stream_fldfilename_lightng = '/glade/p/cesm/cseg/inputdata/atm/datm7/NASA_LIS/
↪clmforc.Li_2012_climo1995-2011.T62.lnfm_c130327.nc'
stream_year_first_lightng = 0001
stream_year_last_lightng = 0001
/
&clm_hydrology1_inparm
/
&clm_soilhydrology_inparm
/
&ch4par_in
fin_use_fsat = .true.
/

```

Adding/removing fields on your primary history file

The primary history files are output monthly, and contain an extensive list of fieldnames, but the list of fieldnames can be added to using `hist_fincl1` or removed from by adding fieldnames to `hist_fexcl1`. A sample user namelist `user_nl_clm` adding few new fields (cosine of solar zenith angle, and solar declination) and excluding a few standard fields is (ground temperature, vegetation temperature, soil temperature and soil water).:

Example 1-3. Example `user_nl_clm` namelist adding and removing fields on primary history file

```

hist_fincl1 = 'COSZEN', 'DECL'
hist_fexcl1 = 'TG', 'TV', 'TSOI', 'H2OSOI'

```

Adding auxiliary history files and changing output frequency

The `hist_fincl2` through `hist_fincl6` set of namelist variables add given history fieldnames to auxiliary history file “streams”, and `hist_fexcl2` through `hist_fexcl6` set of namelist variables remove given history fieldnames from history file auxiliary “streams”. A history “stream” is a set of history files that are produced at a given frequency. By default there is only one stream of monthly data files. To add more streams you add history fieldnames to `hist_fincl2` through `hist_fincl6`. The output frequency and the way averaging is done can be different for each history file stream. By default the primary history files are monthly and any others are daily. You can have up to six active history streams, but you need to activate them in order. So if you activate stream “6” by setting `hist_fincl6`, but if any of `hist_fincl2` through `hist_fincl5` are unset, only the history streams up to the first blank one will be activated.

The frequency of the history file streams is given by the namelist variable `hist_nhtfrq` which is an array of rank six for each history stream. The values of the array `hist_nhtfrq` must be integers, where the following values have the given meaning:

Positive value means the output frequency is the number of model steps between output. *Negative value* means the output frequency is the absolute value in hours given (i.e -1 would mean an hour and -24 would mean a full day). Daily (-24) is the default value for all auxiliary files. *Zero* means the output frequency is monthly. This is the default for the primary history files.

The number of samples on each history file stream is given by the namelist variable `hist_mfilt` which is an array of rank six for each history stream. The values of the array `hist_mfilt` must be positive integers. By default the primary history file stream has one time sample on it (i.e. output is to separate monthly files), and all other streams have thirty time samples on them.

A sample user namelist `user_nl_clm` turning on four extra file streams for output: daily, six-hourly, hourly, and every time-step, leaving the primary history files as monthly, and changing the number of samples on the streams to: yearly (12), thirty, weekly (28), daily (24), and daily (48) is:

Example: user_nl_clm namelist adding auxiliary history files and changing output frequency

```
hist_fincl2 = 'TG', 'TV'  
hist_fincl3 = 'TG', 'TV'  
hist_fincl4 = 'TG', 'TV'  
hist_fincl5 = 'TG', 'TV'  
hist_nhtfrq = 0, -24, -6, -1, 1  
hist_mfilt = 12, 30, 28, 24, 48
```

Removing all history fields

Sometimes for various reasons you want to remove all the history fields either because you want to do testing without any output, or you only want a very small custom list of output fields rather than the default extensive list of fields. By default only the primary history files are active, so technically using `hist_fexcl1` explained in the first example, you could list ALL of the history fields that are output in `hist_fexcl1` and then you wouldn't get any output. However, as the list is very extensive this would be a cumbersome thing to do. So to facilitate this `hist_empty_htapes` allows you to turn off all default output. You can still use `hist_fincl1` to turn your own list of fields on, but you then start from a clean slate. A sample user namelist `user_nl_clm` turning off all history fields and then activating just a few selected fields (ground and vegetation temperatures and absorbed solar radiation) is:

Example 1-5. Example user_nl_clm namelist removing all history fields

```
hist_empty_htapes = .true.  
hist_fincl1 = 'TG', 'TV', 'FSA'
```

Various ways to change history output averaging flags

There are two ways to change the averaging of output history fields. The first is using `hist_avgflag_pertape` which gives a default value for each history stream, the second is when you add fields using `hist_fincl*`, you add an averaging flag to the end of the field name after a colon (for example 'TSOI:X', would output the maximum of TSOI). The types of averaging that can be done are:

- A Average, over the output interval.
- I Instantaneous, output the value at the output interval.
- X Maximum, over the output interval.

- *M* Minimum, over the output interval.

The default averaging depends on the specific fields, but for most fields is an average. A sample user namelist `user_nl_clm` making the monthly output fields all averages (except TSOI for the first two streams and FIRE for the 5th stream), and adding auxiliary file streams for instantaneous (6-hourly), maximum (daily), minimum (daily), and average (daily). For some of the fields we diverge from the per-tape value given and customize to some different type of optimization.

Example: `user_nl_clm` namelist with various ways to average history fields

```
hist_empty_htapes = .true.
hist_fincl1 = 'TSOI:X', 'TG',    'TV',    'FIRE',   'FSR',   'FSH',
              'EFLX_LH_TOT', 'WT'
hist_fincl2 = 'TSOI:X', 'TG',    'TV',    'FIRE',   'FSR',   'FSH',
              'EFLX_LH_TOT', 'WT'
hist_fincl3 = 'TSOI',   'TG:I',  'TV',    'FIRE',   'FSR',   'FSH',
              'EFLX_LH_TOT', 'WT'
hist_fincl4 = 'TSOI',   'TG',    'TV:I',  'FIRE',   'FSR',   'FSH',
              'EFLX_LH_TOT', 'WT'
hist_fincl5 = 'TSOI',   'TG',    'TV',    'FIRE:I', 'FSR',   'FSH',
              'EFLX_LH_TOT', 'WT'
hist_avgflag_pertape = 'A', 'I', 'X', 'M', 'A'
hist_nhtfrq = 0, -6, -24, -24, -24
```

In the example we put the same list of fields on each of the tapes: soil-temperature, ground temperature, vegetation temperature, emitted longwave radiation, reflected solar radiation, sensible heat, total latent-heat, and total water storage. We also modify the soil-temperature for the primary and secondary auxiliary tapes by outputting them for a maximum instead of the prescribed per-tape of average and instantaneous respectively. For the tertiary auxiliary tape we output ground temperature instantaneous instead of as a maximum, and for the fourth auxiliary tape we output vegetation temperature instantaneous instead of as a minimum. Finally, for the fifth auxiliary tapes we output FIRE instantaneously instead of as an average.

Note: We also use `hist_empty_htapes` as in the previous example, so we can list ONLY the fields that we want on the primary history tapes.

Outputting history files as a vector in order to analyze the plant function types within gridcells

By default the output to history files are the grid-cell average of all land-units, and vegetation types within that grid-cell, and output is on the full 2D latitude/longitude grid with ocean masked out. Sometimes it's important to understand how different land-units or vegetation types are acting within a grid-cell. The way to do this is to output history files as a 1D-vector of all land-units and vegetation types. In order to display this, you'll need to do extensive post-processing to make sense of the output. Often you may only be interested in a few points, so once you figure out the 1D indices for the grid-cells of interest, you can easily view that data. 1D vector output can also be useful for single point datasets, since it's then obvious that all data is for the same grid cell.

To do this you use `hist_dov2xy` which is an array of rank six for each history stream. Set it to `.false.` if you want one of the history streams to be a 1D vector. You can also use `hist_type1d_pertape` if you want to average over all the: Plant-Function-Types, columns, land-units, or grid-cells. A sample user namelist `user_nl_clm` leaving the primary monthly files as 2D, and then doing grid-cell (GRID), column (COLS), and no averaging over auxiliary tapes output daily for a single field (ground temperature) is:

Example: user_nl_clm namelist outputting some files in 1D Vector format

```
hist_fincl2 = 'TG'  
hist_fincl3 = 'TG'  
hist_fincl4 = 'TG'  
hist_fincl5 = 'TG'  
hist_fincl6 = 'TG'  
hist_dov2xy = .true., .false., .false., .false.  
hist_type2d_pertape = ' ', 'GRID', 'COLS', ''  
hist_nhtfrq = 0, -24, -24, -24
```

Warning: LAND and COLS are also options to the pertape averaging, but currently there is a bug with them and they fail to work.

Note: Technically the default for hist_nhtfrq is for primary files output monthly and the other auxiliary tapes for daily, so we don't actually have to include hist_nhtfrq, we could use the default for it. Here we specify it for clarity.

Caution: LAND and COLS are also options to the pertape averaging, but currently there is a bug with them and they fail to work.

Visualizing global 1D vector files will take effort. You'll probably want to do some post-processing and possibly just extract out single points of interest to see what is going on. Since, the output is a 1D vector, of only land-points traditional plots won't be helpful. The number of points per grid-cell will also vary for anything, but grid-cell averaging. You'll need to use the output fields pfts1d_ixy, and pfts1d_jxy, to get the mapping of the fields to the global 2D array. pfts1d_itype_veg gives you the PFT number for each PFT. Most likely you'll want to do this analysis in a data processing tool (such as NCL, Matlab, Mathematica, IDL, etcetera that is able to read and process NetCDF data files).

1.2.4 Customizing the DATM namelist

When running “T” compsets with CLM you use the DATM model to give atmospheric forcing data to CLM. There are two ways to customize DATM:

1. **DATM Main Namelist and Stream Namelist gorup** (`datm_in`)
2. **DATM stream files**

The [Data Model Documentation](#) gives the details of all the options for the data models and for DATM specifically. It goes into detail on all namelist items both for DATM and for DATM streams. So here we won't list ALL of the DATM namelist options, nor go into great details about stream files. But, we will talk about a few of the different options that are relevant for running with CLM. All of the options for changing the namelists or stream files is done by editing the `user_nl_datm` file.

Because, they aren't useful for work with CLM we will NOT discuss any of the options for the main DATM namelist. Use the DATM Users Guide at the link above to find details of that. For the streams namelist we will discuss three items:

1. `mapalgo`
2. `taxmode`

3. tintalgo

And for the streams file itself we will discuss:

offset Again everything else (and including the above items) are discussed in the Data Model User's Guide. Of the above the last three: offset, taxmode and tintalgo are all closely related and have to do with the time interpolation of the DATM data.

mapalgo mapalgo sets the spatial interpolation method to go from the DATM input data to the output DATM model grid. The default is bilinear. For CLM1PT we set it to nn to just select the nearest neighbor. This saves time and we also had problems running the interpolation for single-point mode.

taxmode taxmode is the time axis mode. For CLM we usually have it set to cycle which means that once the end of the data is reached it will start over at the beginning. The extend modes is used have it use the last time-step of the forcing data once it reaches the end of forcing data (or use the first time-step before it reaches where the forcing data starts). See the warning below about the extend mode.

Warning: *THE extend OPTION NEEDS TO BE USED WITH CAUTION!* It is only invoked by default for the CLM1PT mode and is only intended for the supported urban datasets to extend the data for a single time-step. If you have the model *run extensively through periods in this mode you will effectively be repeating that last time-step over that entire period*. This means the output of your simulation will be worthless.

offset (in the stream file) offset is the time offset in seconds to give to each stream of data. Normally it is NOT used because the time-stamps for data is set correctly for each stream of data. Note, the offset may NEED to be adjusted depending on the taxmode described above, or it may need to be adjusted to account for data that is time-stamped at the END of an interval rather than the middle or beginning of interval. The offset can be set in the stream file rather than on the stream namelist. For data with a taxmode method of coszen the time-stamp needs to be for the beginning of the interval, while for other data it should be the midpoint. The offset can be used to adjust the time-stamps to get the data to line up correctly.

tintalgo tintalgo is the time interpolation algorithm. For CLM we usually use one of three modes: coszen, nearest, or linear. We use coszen for solar data, nearest for precipitation data, and linear for everything else. If your data is half-hourly or hourly, nearest will work fine for everything. The coszen scaling is useful for longer periods (three hours or more) to try to get the solar to match the cosine of the solar zenith angle over that longer period of time. If you use linear for longer intervals, the solar will cut out at night-time anyway, and the straight line will be a poor approximation of the cosine of the solar zenith angle of actual solar data. nearest likewise would be bad for longer periods where it would be much higher than the actual values.

Note: For coszen the time-stamps of the data should correspond to the beginning of the interval the data is measured for. Either make sure the time-stamps on the datafiles is set this way, or use the offset described above to set it.

Note: For nearest and linear the time-stamps of the data should correspond to the middle of the interval the data is measured for. Either make sure the time-stamps on the datafiles is set this way, or use the offset described above to set it.

In the sections below we go over each of the relevant DATM_MODE options and what the above DATM settings are for each. This gives you examples of actual usage for the settings. We also describe in what ways you might want to customize them for your own case.

CLMCRUNCEP mode and it's DATM settings

In CLMCRUNCEP mode the CRUNCEP dataset is used and all of it's data is on a 6-hourly interval. Like CLM_QIAN the dataset is divided into those three data streams: solar, precipitation, and everything else (temperature, pressure, humidity and wind). The time-stamps of the data were also adjusted so that they are the beginning of the interval for solar, and the middle for the other two. Because, of this the `offset` is set to zero, and the `tintalgo` is: `coszen`, `nearest`, and `linear` for the solar, precipitation and other data respectively. `taxmode` is set to `cycle` and `mapalgo` is set to `bilinear` so that the data is spatially interpolated from the input exact half degree grid to the grid the atmosphere model is being run at (to run at this same model resolution use the 360x720cru_360x720cru resolution).

Note: The “everything else” data stream (of temperature, pressure, humidity and wind) also includes the data for longwave downward forcing as well. Our simulations showed sensitivity to this field, so we backed off in using it, and let DATM calculate longwave down from the other fields.

For more information on CRUNCEP forcing see <http://dods.extra.cea.fr/data/p529viov/cruncep/>.

CLM_QIAN mode and it's DATM settings

In CLM_QIAN mode the Qian dataset is used which has 6-hourly solar and precipitation data, and 3-hourly for everything else. The dataset is divided into those three data streams: solar, precipitation, and everything else (temperature, pressure, humidity and wind). The time-stamps of the data were also adjusted so that they are the beginning of the interval for solar, and the middle for the other two. Because, of this the `offset` is set to zero, and the `tintalgo` is: `coszen`, `nearest`, and `linear` for the solar, precipitation and other data respectively. `taxmode` is set to `cycle` and `mapalgo` is set to `bilinear` so that the data is spatially interpolated from the input T62 grid to the grid the atmosphere model is being run at.

Normally you wouldn't customize the CLM_QIAN settings, but you might replicate it's use for your own global data that had similar temporal characteristics.

CLM1PT mode and it's DATM settings

In CLM1PT mode the model is assumed to have half-hourly or hourly data for a single-point. For the supported datasets that is exactly what it has. But, if you add your own data you may need to make adjustments accordingly. Using the `CLM_USRDATA_NAME` resolution you can easily extend this mode for your own datasets that may be regional or even global and could be at different temporal frequencies. If you do so you'll need to make adjustments to your DATM settings. The dataset has all data in a single stream file. The time-stamps of the data were also adjusted so that they are at the middle of the interval. Because, of this the `offset` is set to zero, and the `tintalgo` is set to `nearest`. `taxmode` is set to `extend` and `mapalgo` is set to `nn` so that simply the nearest point is used.

If you are using your own data for this mode and it's not at least hourly you'll want to adjust the DATM settings for it. If the data is three or six hourly, you'll need to divide it up into separate streams like in CLM_QIAN mode which will require fairly extensive changes to the DATM namelist and streams files. For an example of doing this see Example 5-8.

CPLHIST3HrWx mode and it's DATM settings

In CPLHIST3HrWx mode the model is assumed to have 3-hourly for a global grid from a previous CESM simulation. Like CLM_QIAN mode the data is divided into three streams: one for precipitation, one for solar, and one for everything else. The time-stamps for Coupler history files for CESM is at the end of the interval, so the `offset` needs to be set in order to adjust the time-stamps to what it needs to be for the `tintalgo` settings. For precipitation `taxmode` is set to `nearest` so the `offset` is set to -5400 seconds so that the ending time-step is adjusted by an hour and

half to the middle of the interval. For solar `taxmode` is set to `coszen` so the offset is set to -10800 seconds so that the ending time-step is adjust by three hours to the beginning of the interval. For everything else `taxmode` is set to `linear` so the offset is set to -5400 seconds so that the ending time-step is adjusted by an hour and half to the middle of the interval. For an example of such a case see the [Section called Running with MOAR data as atmospheric forcing to spinup the model in Chapter 4](#).

Normally you wouldn't modify the DATM settings for this mode. However, if you had data at a different frequency than 3-hours you would need to modify the `offset` and possibly the `taxmode`. The other two things that you might modify would be the path to the data or the domain file for the resolution (which is currently hardwired to f09). For data at a different input resolution you would need to change the domain file in the streams file to use a domain file to the resolution that the data comes in on.

1.3 Using CLM tools

1.3.1 What are the CLM tools

There are several tools provided with CLM that allow you to create your own input datasets at resolutions you choose, or to interpolate initial conditions to a different resolution, or used to compare CLM history files between different cases. The tools are all available in the `models/lnd/clm/tools` directory. Most of the tools are FORTRAN stand-alone programs in their own directory, but there is also a suite of NCL scripts in the `shared/ncl_scripts` directory, and some of the tools are scripts that may also call the ESMF regridding program. Some of the NCL scripts are very specialized and not meant for general use, and we won't document them here. They still contain documentation in the script itself and the README file in the tools directory.

The tools are divided into three directories for three categories: `clm4_0`, `clm4_5`, and `shared`. The first two are of course for tools that are designed to work with either the CLM4.0 or CLM4.5 versions of the model. The last one are shared utilities that can be used by either, or have a “-phys” option so you can specify which version you want to use.

The list of generally important scripts and programs are as follows.

1. `tools/cprnc` (relative to top level directory) to compare NetCDF files with a time axis.
2. `shared/mkmapgrids` to create SCRIP grid data files from old CLM format grid files that can then be used to create new CLM datasets (deprecated). There is also a NCL script (`shared/mkmapgrids/mkscripgrid.ncl`) to create SCRIP grid files for regular latitude/longitude grids.
3. `shared/mkmapdata` to create SCRIP mapping data file from SCRIP grid files (uses ESMF).
4. `shared/gen_domain` to create a domain file for datm from a mapping file. The domain file is then used by BOTH datm AND CLM to define the grid and land-mask.
5. `mksurfdatal_map` to create surface datasets from grid datasets (`clm4_0` and `clm4_5` versions).
6. `shared/mkprodata_map` to interpolate output unstructured grids (such as the CAM HOMME dy-core “ne” grids like `ne30np4`) into a 2D regular lat/long grid format that can be plotted easily. Can be used by either `clm4_0` or `clm4_5`.

In the sections to come we will go into detailed description of how to use each of these tools in turn. First, however we will discuss the common environment variables and options that are used by all of the FORTRAN tools. Second, we go over the outline of the entire file creation process for all input files needed by CLM for a new resolution, then we turn to each tool. In the last section we will discuss how to customize files for particular observational sites.

The tools run either one of two ways, with a namelist to provide options, or with command line arguments (and NOT both). `gen_domain` and `cprnc` run with command line arguments, and the other tools run with namelists.

In the following sections, we will outline how to make these files available for build-namelist so that you can easily create simulations that include them. In the chapter on single-point and regional datasets we also give an alternative way to enter new datasets without having to edit files.

Running FORTRAN tools with namelists

mksurfdata_map and **mkmapgrids** run with namelists that are read from standard input. Hence, you create a namelist and then run them by redirecting the namelist file into standard input as follows:

```
./program < namelist
```

For programs with namelists there is at least one sample namelist with the name “program”.namelist (i.e. mksurfdata_map.namelist for the **mksurfdata_map** program). There may also be other sample namelists that end in a different name besides “namelist”. Namelists that you create should be similar to the example namelist. The namelist values are also documented along with the other namelists in the:

```
models/lnd/clm/bld/namelist_files/namelist_definition.xml`` file  
and default values in the:  
models/lnd/clm/bld/namelist_files/namelist_defaults_clm_tools.xml`` file.
```

Running FORTRAN tools with command line options

gen_domain, and **cprnc** run with command line arguments. The detailed sections below will give you more information on the command line arguments specific to each tool. Also running the tool without any arguments will give you a general synopsis on how to run the tool.

Running FORTRAN tools built with SMP=TRUE

When you enable **SMP=TRUE** on your build of one of the tools that make use of it, you are using OpenMP for shared memory parallelism (SMP). In SMP loops are run in parallel with different threads run on different processors all of which access the same memory (called on-node). Thus you can only usefully run up to the number of processors that are available on a single-node of the machine you are running on. For example, on the NCAR machine yellowstone there are 16 processors per node, but the SMT hardware on the machine allows you to submit twice as many threads or 32 threads.

Using NCL

In the tools directory `models/lnd/clm/tools/shared/ncl_scripts` and in a few other locations there are scripts that use NCAR Command Language (NCL). Unlike the FORTRAN tools, you will need to get a copy of NCL in order to use them. You also won’t have to build an executable in order to use them, hence no Makefile is provided. NCL is provided for free download as either binaries or source code from: <http://www.ncl.ucar.edu/>. The NCL website also contains documentation on NCL and its use. These scripts are stand-alone and at most use environment variables to control how to use them. In some cases there are perl scripts with command line arguments that call the NCL scripts to control what they do.

1.3.2 Building the CLM tools

The FORTRAN tools all have similar makefiles, and similar options for building. All of the Makefiles use GNU Make extensions and thus require that you use GNU make to use them. They also auto detect the type of platform you are on, using “uname -s” and set the compiler, compiler flags and such accordingly. There are also environment variables that can be set to set things that must be customized. All the tools use NetCDF and hence require the path to the NetCDF libraries and include files. On some platforms (such as Linux) multiple compilers can be used, and hence there are env variables that can be set to change the FORTRAN and/or “C” compilers used. The tools other than **cprnc** also allow finer control, by also allowing the user to add compiler flags they choose, for both FORTRAN and “C”, as well as picking the compiler, linker and and add linker options. Finally the tools other than **cprnc** allow you to turn

optimization on (which is off by default but on for the **mksurfdata_map** and **interpinic** programs) with the OPT flag so that the tool will run faster. To get even faster performance, the **interpinic**, program allows you to also use the SMP to turn on multiple shared memory processors. When SMP=TRUE you set the number of threads used by the program with the OMP_NUM_THREADS environment variable.

Options used by all: **cprnc**, **interpinic**, and **mksurfdata_map**

- LIB_NETCDF – sets the location of the NetCDF library.
- INC_NETCDF – sets the location of the NetCDF include files.
- USER_FC – sets the name of the FORTRAN compiler.

Options used by: **interpinic**, **mkproCDATA_map**, **mkmapgrids**, and **mksurfdata_map**

- MOD_NETCDF – sets the location of the NetCDF FORTRAN module.
- USER_LINKER – sets the name of the linker to use.
- USER_CPPDEFS – adds any CPP defines to use.
- USER_CFLAGS – add any “C” compiler flags to use.
- USER_FFLAGS – add any FORTRAN compiler flags to use.
- USER_LDFLAGS – add any linker flags to use.
- USER_CC – sets the name of the “C” compiler to use.
- OPT – set to TRUE to compile the code optimized (TRUE or FALSE)
- SMP – set to TRUE to turn on shared memory parallelism (i.e. OpenMP) (TRUE or FALSE)
- Filepath – list of directories to build source code from.
- Srcfiles – list of source code filenames to build executable from.
- Makefile – customized makefile options for this particular tool.
- mkDepends – figure out dependencies between source files, so make can compile in order..
- Makefile.common – General tool Makefile that should be the same between all tools.

Options used only by **cprnc**:

- EXEDIR – sets the location where the executable will be built.
- VPATH – colon delimited path list to find the source files.

More details on each environment variable.

LIB_NETCDF This variable sets the path to the NetCDF library file (`libnetcdf.a`). If not set it defaults to `/usr/local/lib`. In order to use the tools you need to build the NetCDF library and be able to link to it. In order to build the model with a particular compiler you may have to compile the NetCDF library with the same compiler (or at least a compatible one).

INC_NETCDF This variable sets the path to the NetCDF include directory (in order to find the include file `netcdf.h`). if not set it defaults to `/usr/local/include`.

MOD_NETCDF This variable sets the path to the NetCDF module directory (in order to find the NetCDF FORTRAN-90 module file when NetCDF is used with a FORTRAN-90 **use statement**). When not set it defaults to the **LIN_NETCDF** value.

USER_FC This variable sets the command name to the FORTRAN-90 compiler to use when compiling the tool. The default compiler to use depends on the platform. And for example, on the AIX platform this variable is NOT used

USER_LINKER This variable sets the command name to the linker to use when linking the object files from the compiler together to build the executable. By default this is set to the value of the FORTRAN-90 compiler used to compile the source code.

USER_CPPDEFS This variable adds additional optional values to define for the C preprocessor. Normally, there is no reason to do this as there are very few CPP tokens in the CLM tools. However, if you modify the tools there may be a reason to define new CPP tokens.

USER_CC This variable sets the command name to the “C” compiler to use when compiling the tool. The default compiler to use depends on the platform. And for example, on the AIX platform this variable is NOT used

USER_CFLAGS This variable adds additional compiler options for the “C” compiler to use when compiling the tool. By default the compiler options are picked according to the platform and compiler that will be used.

USER_FFLAGS This variable adds additional compiler options for the FORTRAN-90 compiler to use when compiling the tool. By default the compiler options are picked according to the platform and compiler that will be used.

USER_LDFLAGS This variable adds additional options to the linker that will be used when linking the object files into the executable. By default the linker options are picked according to the platform and compiler that is used.

SMP This variable flags if shared memory parallelism (using OpenMP) should be used when compiling the tool. It can be set to either TRUE or FALSE, by default it is set to FALSE, so shared memory parallelism is NOT used. When set to TRUE you can set the number of threads by using the OMP_NUM_THREADS environment variable. Normally, the most you would set this to would be to the number of on-node CPU processors. Turning this on should make the tool run much faster.

Warning: Note, that depending on the compiler answers may be different when SMP is activated.

OPT This variable flags if compiler optimization should be used when compiling the tool. It can be set to either TRUE or FALSE, by default it is set to FALSE for **mkmapgrids** and TRUE for **mksurfdatal_map**, **mkprocdatal_map** and **interpinic**. Turning this on should make the tool run much faster.

Warning: Note, you should expect that answers will be different when OPT is activated.

Filepath All of the tools are stand-alone and don’t need any outside code to operate. The Filepath is the list of directories needed to compile and hence is always simply “.” the current directory. Several tools use copies of code outside their directory that is in the CESM distribution (either csm_share code or CLM source code).

Srcfiles The Srcfiles lists the filenames of the source code to use when building the tool.

Makefile The Makefile is the custom GNU Makefile for this particular tool. It will customize the EXENAME and the optimization settings for this particular tool.

Makefile.common The Makefile.common is the copy of the general GNU Makefile for all the CLM tools. This file should be identical between the different tools. This file has different sections of compiler options for different Operating Systems and compilers.

mkDepends The mkDepends is the copy of the perl script used by the Makefile.common to figure out the dependencies between the source files so that it can compile in the necessary order. This file should be identical between the different tools.

EXEDIR The cprnc tool uses this variable to set the location of where the executable will be built. The default is the current directory.

VPATH The cprnc tool uses this variable to set the colon delimited pathnames of where the source code exists. The default is the current directory.

Note: There are several files that are copies of the original files from either models“/lnd/clm/src/util_share“, models/csm_share/shr, or copies from other tool directories. By having copies the tools can all be made stand-alone, but any changes to the originals will have to be put into the tool directories as well.

The *README.filecopies* (which can be found in models/lnd/clm/tools) is repeated here.

models/lnd/clm/tools/README.filecopies	Jun/04/2013
--	-------------

There are several files that are copies of the original files **from either** models/lnd/clm/src/main, models/csm_share/shr, models/csm_share/unit_testers, **or** copies **from other** tool directories. By having copies the tools can **all** be made stand-alone, but **any** changes to the originals will have to be put into the tool directories **as** well.

I. Files that are IDENTICAL:

1. csm_share files copied that should be identical to models/csm_share/shr:

```
shr_const_mod.F90
shr_log_mod.F90
shr_timer_mod.F90
```

2. csm_share files copied that should be identical to models/csm_share/unit_testers:

```
test_mod.F90
```

3. clm/src files copied that should be identical to models/lnd/clm/src/util_share:

```
nanMod.F90
```

II. Files **with** differences

1. csm_share files copied **with** differences:

```
shr_kind_mod.F90 --- SHR_KIND_CXX is new
shr_sys_mod.F90 ---- Remove mpi abort and reference to shr_mpi_mod.F90.
shr_infnan_mod.F90 - Earlier version
shr_string_mod.F90 - Earlier version
shr_file_mod.F90 --- mkprocdat_map version is stripped down
clm_varctl.F90 ----- Earlier version
```

2. clm/src files **with** differences:

```
fileutils.F90 --- Remove use of masterproc and spmdMod and endrun in abortutils.
```

4. Files **in** mkmappgrids

```
domainMod.F90 ---- Highly customized based off an earlier version of clm code.
                    Remove use of abortutils, spmdMod. clm version uses latlon
                    this version uses domain in names. Distributed memory
                    parallelism is removed.
```

5. Files **in** mksurfdata_map

```
mkvarpar.F90 --- clm4_0 and clm4_5 versions are different and different from  

main clm versions.
```

1.3.3 Creating input for surface dataset generation

1. Generating SCRIP grid files

The utility `mkmapdata.sh` requires SCRIP format input files to describe the input and output grids that maps are generated for. CLM provides a utility, `mkmapgrids` that generates those files. The program converts old formats of CAM or CLM grid files to SCRIP grid format. There is also a NCL script (`mkscripgrid.ncl`) to create regular latitude longitude regional or single-point grids at the resolution the user desires.

SCRIP grid files for all the standard model resolutions and the raw surface datasets have already been done and the files are in the XML database. Hence, this step doesn't need to be done – EXCEPT WHEN YOU ARE CREATING YOUR OWN GRIDS. If you have a CLM grid or CAM file from previous versions and you want to convert it you can use `mkmapgrids`.

Using `mknoocnmap.pl` to create grid and maps for single-point regional grids

If you want to create a regular latitude/longitude single-point or regional grid, we suggest you use `mknoocnmap.pl` in `models/lnd/clm/tools/shared/mkmapdata` which will create both the SCRIP grid file you need (using `models/lnd/clm/tools/shared/mkmapgrids/mkscripgrid.ncl` AND an identity mapping file assuming there is NO ocean in your grid domain. If you HAVE ocean in your domain you could modify the mask in the SCRIP grid file for ocean, and then use `ESMF_RegridWeightGen` to create the mapping file, and `gen_domain` to create the domain file. Like other tools, `shared/mkmapdata/mknoocnmap.pl` has a help option with the following:

```
SYNOPSIS
  mknoocnmap.pl [options]      Gets map and grid files for a single land-only point.

REQUIRED OPTIONS
  -centerpoint [or -p] <lat,lon> Center latitude,longitude of the grid to create.
  -name [-or -n] <name>        Name to use to describe point

OPTIONS
  -dx <number>                Size of total grid in degrees in longitude
  direction                  (default is 0.1)
  -dy <number>                Size of total grid in degrees in latitude
  direction                  (default is 0.1)
  -silent [or -s]             Make output silent
  -help [or -h]               Print usage to STDOUT.
  -verbose [or -v]             Make output more verbose.
  -nx <number>                Number of longitudes (default is 1)
  -ny <number>                Number of latitudes (default is 1)
```

See Figure 2-5 for a visual representation of this process.

2. Creating mapping files for `mksurfdatal_map`

`mkmapdata.sh` uses the above SCRIP grid input files to create SCRIP mapping data files (uses ESMF).

The bash shell script `models/lnd/clm/tools/shared/mkmapgrids/mkmapdata.sh` uses **ESMF_RegridWeightGen** to create a list of maps from the raw datasets that are input to **mksurfdata_map**. Each dataset that has a different grid, or land-mask needs a different mapping file for it, but many different raw datasets share the same grid/land-mask as other files. Hence, there doesn't need to be a different mapping file for EACH raw dataset – just for each DIFFERENT raw dataset. See Figure 2-3 for a visual representation of how this works. The bash script figures out which mapping files it needs to create and then runs **ESMF_RegridWeightGen** for each one. You can then either enter the datasets into the XML database (see Chapter 3 or leave the files in place, and use the “-res usrspec -usr_gname -usr_gdate” options to **mksurfdata_map** (see the Section called Running `mksurfdata.pl` below). Use the “-phys” option to specify if you are creating mapping files for clm4_0 or clm4_5 (the list of raw datafiles is somewhat different between the two). `mkmapdata.sh` has a help option with the following

```
../../../../tools/shared/mkmapdata/mkmapdata.sh

*****
usage on yellowstone:
./mkmapdata.sh

valid arguments:
[-f|--gridfile <gridname>]
    Full pathname of model SCRIP grid file to use
    This variable should be set if this is not a supported grid
    This variable will override the automatic generation of the
    filename generated from the -res argument
    the filename is generated ASSUMING that this is a supported
    grid that has entries in the file namelist_defaults_clm.xml
    the -r|--res argument MUST be specified if this argument is specified
[-r|--res <res>]
    Model output resolution (default is 10x15)
[-t|--gridtype <type>]
    Model output grid type
    supported values are [regional,global], (default is global)
[-p|--phys <CLM-version>]
    Whether to generate mapping files for clm4_0 or clm4_5
    supported values are [clm4_0,clm4_5], (default is clm4_5)
[-b|--batch]
    Toggles batch mode usage.
If you want to run in batch mode
    you need to have a separate batch script for a supported machine
    that calls this script interactively - you cannot submit this
    script directory to the batch system
[-l|--list]
    List mapping files required (use check_input_data to get them)
    also writes data to clm.input_data_list
[-d|--debug]
    Toggles debug-only (don't actually run mkmapdata just echo what would happen)
[-h|--help]
    Displays this help message
[-v|--verbose]
    Toggle verbose usage -- log more information on what is happening

You can also set the following env variables:
ESMFBIN_PATH - Path to ESMF binaries
                (default is /contrib/esmf-5.3.0-64-0/bin)
CSMDATA ----- Path to CESM input data
                (default is /glade/p/cesm/cseg/inputdata)
MPIEXEC ----- Name of mpirun executable
                (default is mpirun.lsf)
REGRID_PROC -- Number of MPI processors to use
```

```
(default is 8)

**pass environment variables by preceding above commands
with 'env var1=setting var2=setting'
*****
```

Warning: Make sure you specify with the “-phys” option if you are creating files for CLM4.0! The default is CLM4.5.

Figure 2-3. Details of running `mkmapdata.sh`

Insert figure 2-3

Each of the raw datasets for `mksurfdata_map` needs a mapping file to map from the output grid you are running on to the grid and land-mask for that dataset. This is what `mkmapdata.sh` does. To create the mapping files you need a SCRIP grid file to correspond with each resolution and land mask that you have a raw data file in `mksurfdata_map`. Some raw datasets share the same grid and land mask – hence they can share the same SCRIP grid file. The output maps created here go into `mksurfdata_map` see [Figure 2-6](#).

1.3.4 Creating Surface Datasets

When just creating a replacement file for an existing one, the relevant tool should be used directly to create the file. When you are creating a set of files for a new resolution there are some dependencies between the tools that you need to keep in mind when creating them. The main dependency is that you MUST create a SCRIP grid file first as the SCRIP grid dataset is then input into the other tools. Also look at [Table 3-1](#) which gives information on the files required and when. [Figure 2-1](#) shows an overview of the general data-flow for creation of the fsurdat datasets.

Figure 2-1. Data Flow for Creation of Surface Datasets from Raw SCRIP Grid Files

Insert figure 2-1

Starting from a SCRIP grid file that describes the grid you will run the model on, you first run `mkmapdata.sh` to create a list of mapping files. See [Figure 2-3](#) for a more detailed view of how `mkmapdata.sh` works. The mapping files tell `mksurfdata_map` how to map between the output grid and the raw datasets that it uses as input. The output of `mksurfdata_map` is a surface dataset that you then use for running the model. See [Figure 2-6](#) for a more detailed view of how `mksurfdata_map` works.

[Figure 2-2](#) is the legend for this figure ([Figure 2-1](#)) and other figures in this chapter ([Figure 2-4](#), [Figure 2-5](#), and [Figure 2-6](#)). [Figure 2-2](#). Legend for Data Flow Figures Insert figure 2-2

Green arrows define the input to a program, while red arrows define the output. Cylinders define files that are either created by a program or used as input for a program. Boxes are programs.

You start with a description of a SCRIP grid file for your output grid file and then create mapping files from the raw datasets to it. Once, the mapping files are created `mksurfdata_map` is run to create the surface dataset to run the model.

Creating a Complete Set of Files for Input to CLM

1. Create SCRIP grid datasets (if NOT already done)

First you need to create a descriptor file for your grid, that includes the locations of cell centers and cell corners. There is also a “mask” field, but in this case the mask is set to one everywhere (i.e. all of the masks for the output model grid are “nomask”). An example SCRIP grid file is: \$CSM-DATA/lnd/clm/mappingdata/grids/SCRIPgrid_10x15_nomask_c110308.nc. The mkmapgrids and mkscripgrid.ncl NCL script in the models/lnd/clm/tools/shared/mkmapgrids directory can help you with this. SCRIP grid files for all the standard CLM grids are already created for you. See the Section called Creating an output SCRIP grid file at a resolution to run the model on for more information on this.

2. Create domain dataset (if NOT already done)

Next use gen_domain to create a domain file for use by DATM and CLM. This is required, unless a domain file was already created. See the Section called Creating a domain file for CLM and DATM for more information on this.

3. Create mapping files for mksurfdata_map (if NOT already done)

Create mapping files for mksurfdata_map with mkmapdata.sh in models/lnd/clm/tools/shared/mkmapdata. See the Section called Creating mapping files that mksurfdata_map will use for more information on this.

4. Create surface datasets

Next use mksurfdata_map to create a surface dataset, using the mapping datasets created on the previous step as input. There is a version for either clm4_0 or clm4_5 for this program. See the Section called Using mksurfdata_map to create surface datasets from grid datasets for more information on this.

5. Create some sort of initial condition dataset

You then need to do one of the following three options to have an initial dataset to start from.

- (a) Use spinup-procedures to create initial condition datasets

The first option is to do the spinup procedures from arbitrary initial conditions to get good initial datasets. This is the most robust method to use. See the Section called Spinning up the Satellite Phenology Model (CLMSP spinup) in Chapter 4, the Section called Spinning up the CLM4.0 biogeochemistry Carbon-Nitrogen Model (CN spinup) in Chapter 4, or the Section called Spinning up the CLM4.0 Carbon-Nitrogen Dynamic Global Vegetation Model (CNDV spinup) in Chapter 4 for more information on this.

- (b) Use interpinic to interpolate existing initial condition datasets

The next option is to interpolate from spunup datasets at a different resolution, using interpinic. There is a version for either clm4_0 or clm4_5 for this program. See the Section called Using interpinic to interpolate initial conditions to different resolutions for more information on this.

- (c) Start up from arbitrary initial conditions

The last alternative is to run from arbitrary initial conditions without using any spun-up datasets. This is inappropriate when using CLM4.5-BGC or CLMCN (bgc=cn or cndv) as it takes a long time to spinup Carbon pools.

Warning: This is NOT recommended as many fields in CLM take a long time to equilibrate.

6. Enter the new datasets into the build-namelist XML database The last optional thing to do is to enter the new datasets into the build-namelist XML database. See Chapter 3 for more information on doing this. This is optional because the user may enter these files into their namelists manually. The advantage of entering them into the database is so that they automatically come up when you create new cases.

The models/lnd/clm/tools/README goes through the complete process for creating input files needed to run CLM. We repeat that file here:

models/lnd/clm/tools/README

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CLM tools for analysis of CLM history files -- or for creation or modification of CLM input files.

I. General directory structure:

```
clm4_0
    mksurfdata_map --- Create surface datasets.
    interpinic ----- Interpolate initial datasets to a different resolution.
                        (has optimized and OMP options)

clm4_5
    mksurfdata_map --- Create surface datasets.
    interpinic ----- Interpolate initial datasets to a different resolution.
                        (has optimized and OMP options)

shared
    mkmapgrids ----- Create SCRIP grid files needed by mkmapdata
                        [input is CLM grid files]
                        (deprecated)
    mkmapdata ----- Create SCRIP mapping data from SCRIP grid files (uses ESMF)
    gen_domain ----- Create data model domain datasets from SCRIP mapping datasets.
                        (also in the top level mapping directory [.../.../.../.../tools/
→mapping])
    mkprocdata_map --- Convert output unstructured grids into a 2D format that
                        can be plotted easily
    ncl_scripts ----- NCL post or pre processing scripts.
```

Note that there are different versions of mksurfdata_map and interpinic for CLM4.0 vs. CLM4.5. Other tools are shared between the two model versions.

However, note that mkmapdata makes mapping files for CLM4.5 by default; to make mapping files for CLM4.0, run the tool with the option:

```
-p clm4_0
```

II. Notes on building/running for each of the above tools:

Each tool that has FORTRAN source code has the following files:

```
README ----- Specific help for using the specific tool and help on_
→specific
                    files in that directory.
src/Filepath ----- List of directories needed to build the tool
                    (some files in ../src directories are required).
src/Makefile ----- Customization of the make for the particular tool in_
→question
                    src/Makefile.common - General GNU Makefile for creating FORTRAN tools
                    (these are identical between tools).
src/Srcfiles ----- List of source files that are needed.
src/Mkdepends ----- Dependency generator program
```

mkmapdata and ncl_scripts only contain scripts so don't have the above build files.

Most tools have copies of files from other directories -- see the README.filecopies file for more information on this.

Tools may also have files with the directory name followed by: namelist, or [runoptions](#).

```
<directory>.namelist ----- Namelist to create a global file.  
<directory>.runoptions ---- Command line options to use the given tool.
```

These files are also used by the test scripts to test the tools (see the README.testing) file.

NOTE: Be sure to change the path of the datasets references by these namelists to point to where you have exported your CESM inputdata datasets.

To build:

```
cd <directory>  
setenv INC_NETCDF <path-to-NetCDF-include-files>  
setenv LIB_NETCDF <path-to-NetCDF-library-files>  
gmake
```

The process will create a file called "Depends" which has the dependencies for the build of each file on other files.

By default some codes may be compiled non-optimized so that you can use the debugger, and with bounds-checking, and float trapping on. To speed up do the following...

```
gmake OPT=TRUE (by default already on for interpinic and mksurfdata_map)
```

Also some of the tools allow for OpenMP shared memory parallelism (such as interpinic) with

```
gmake SMP=TRUE
```

To run a program with a namelist:

```
./program < namelist
```

To get help on running a program with command line options (e.g., interpinic):

```
./program
```

To run a program built with SMP=TRUE:

```
setenv OMP_NUM_THREADS=<number_of_threads_to_use>
```

run normally as above

III. Process sequence to create input datasets needed to run CLM

NOTE: The following assumes you want to create files for CLM4.5. If you want to use CLM4.0, you will need to do the following:

- In the following commands, change references to the clm4_5 directory to clm4_0
- Add the option '-p clm4_0' to the mkmapdata.sh command.

1.) Create SCRIP grid files (if needed)

a.) For standard resolutions these files will already be created. (done)

b.) To create regular lat-lon regional/single-point grids run `mknoocnmap.pl`

This will create both SCRIP grid files and a mapping file that will be valid if the region includes NO ocean whatsoever (so you can skip step 2). You can also use this script to create SCRIP grid files for a region (or even a global grid) that DOES include ocean if you use step 2 to create mapping files for it (simply discard the non-ocean map created by this script).

Example, for single-point over Boulder Colorado.

```
cd shared/mkmapdata
./mknoocnmap.pl -p 40,255 -n 1x1_boulderCO
```

c.) General case

You'll need to convert or create SCRIP grid files on your own (using scripts or other tools) for the general case where you have an unstructured grid, or a grid that is not regular in latitude and longitude.

example format

```
=====
netcdf fv1.9x2.5_090205 {
dimensions:
    grid_size = 13824 ;
    grid_corners = 4 ;
    grid_rank = 2 ;
variables:
    double grid_center_lat(grid_size) ;
        grid_center_lat:units = "degrees" ;
    double grid_center_lon(grid_size) ;
        grid_center_lon:units = "degrees" ;
    double grid_corner_lat(grid_size, grid_corners) ;
        grid_corner_lat:units = "degrees" ;
    double grid_corner_lon(grid_size, grid_corners) ;
        grid_corner_lon:units = "degrees" ;
    int grid_dims(grid_rank) ;
    int grid_imask(grid_size) ;
        grid_imask:units = "unitless" ;
```

2.) Create ocean to atmosphere mapping file (if needed)

a.) Standard resolutions (done)

If this is a standard resolution with a standard ocean resolution -- this step is already done, the files already exist.

b.) Region without Ocean (done in step 1.b)

IF YOU RAN `mknoocnmap.pl` FOR A REGION WITHOUT OCEAN THIS STEP IS ALREADY DONE.

c.) New atmosphere or ocean resolution

If the region DOES include ocean, use `gen_domain` to create a mapping file for it.

Example:

```
cd ../../../../tools/mapping/gen_domain_files/src
./gen_domain -m $MAPFILE -o $OCNGRIDNAME -l $ATMGRIDNAME
```

3.) Add SCRIP grid file(s) created in (1) into XML database in CLM (optional)

See the "Adding New Resolutions or New Files to the build-namelist Database" Chapter in the CLM User's Guide

<http://www.cesm.ucar.edu/models/cesm1.0/clm/models/lnd/clm/doc/UsersGuide/book1.html>

If you don't do this step, you'll need to specify the file to mkmapdata in step (3) using the "-f" option.

4.) Create mapping files for use by mksurfdata_map with mkmapdata
(See mkmapdata/README for more help on doing this)

- this step uses the results of (1) that were entered into the XML database by step (3). If you don't enter datasets in, you need to specify the SCRIP grid file using the "-f" option to mkmapdata.sh.
- note that mkmapdata generates maps for CLM4.5 by default; to generate mapping files for CLM4.0, add the option '-p clm4_0'

Example: to generate all necessary mapping files for the ne30np4 grid

```
cd shared/mkmapdata
./mkmapdata.sh -r ne30np4
```

5.) Add mapping file(s) created in step (4) into XML database in CLM (optional)

See notes on doing this in step (3) above.

Edit ../bld/namelist_files/namelist_defaults_clm.xml to incorporate new mapping files.

If you don't do this step, you'll need to specify the grid resolution name and file creation dates to mksurfdata_map in step (5) below.

6.) Convert map of ocean to atm for use by DATM and CLM with gen_domain
(See tools/mapping/README for more help on doing this)

- gen_domain uses the map from step (2) (or previously created CESM maps)

Example:

```
cd ../../../../tools/mapping/gen_domain_files/src
gmake
cd ..
setenv CDATE      090206
setenv OCNGRIDNAME gx1v6
setenv ATMGRIDNAME fv1.9x2.5
setenv MAPFILE ${CSMDATA}/cpl/cpl16/map_${OCNGRIDNAME}_to_${ATMGRIDNAME}_aave_da_$
→{CDATE}.nc
./gen_domain -m $MAPFILE -o $OCNGRIDNAME -l $ATMGRIDNAME
```

Normally for I compsets running CLM only you will discard the ocean domain file, and only use the atmosphere domain file for datm and as the fatmlndfrc

file for CLM. Output domain files will be named according to the input OCN/LND gridnames.

- 7.) Create surface datasets with mksurfdata_map
(See mksurfdata_map/README for more help on doing this)

- Run clm4_5/mksurfdata_map/mksurfdata.pl
- This step uses the results of step (4) entered into the XML database in step (5).
- If datasets were NOT entered into the XML database, set the resolution to "usrspec" and use the "-usr_gname", and "-usr_gdate" options.

Example: for 0.9x1.25 resolution

```
cd clm4_5/mksurfdata_map/src
gmake
cd ..
./mksurfdata.pl -r 0.9x1.25
```

NOTE that surface dataset will be used by default for fatmgrd - and it will contain the lat,lon,edges and area values for the atm grid - ASSUMING that the atm and land grid are the same

- 8.) Interpolate initial conditions using interpinic (optional)

(See interpinic/README for more help on doing this)

IMPORTANT NOTE on interpinic!!!!: BE SURE TO USE NetCDF4.3 WHEN BUILDING!

If your template file was written using pnetcdf -- interpinic will corrupt the resulting file and make it unusable!

- 9.) Add new files to XML data or using user_nl_clm (optional)

See notes on doing this in step (3) above.

IV. Example of creating single-point datasets without entering into XML database.

Here we apply the process described in III. for a single-point dataset where we don't enter the datasets into the XML database (thus skipping steps 3, 5 and 9), but use the needed command line options to specify where the files are. This also skips step (2) since step 1 creates the needed mapping file. We also skip step (8) and do NOT create a finidat file.

- 0.) Set name of grid to use and the creation date to be used later...
setenv GRIDNAME 1x1_boulderCO
setenv CDATE `date +%y%m%d`
- 1.) SCRIP grid and atm to ocn mapping file
cd shared/mkmapdata
.mknoocnmap.pl -p 40,255 -n \$GRIDNAME
Set pointer to MAPFILE that will be used in step (6)
setenv MAPFILE `pwd`/map_\${GRIDNAME}_nocean_to_\${GRIDNAME}_nomask_aave_da_\$
→{CDATE}.nc
cd ..
- 2.) skip
- 3.) skip
- 4.) Mapping files needed for mksurfdata_map
cd shared/mkmapdata
setenv GRIDFILE ../mkmapgrids/SCRIPgrid_\${GRIDNAME}_nomask_\${CDATE}.nc
.mkmapdata.sh -r \$GRIDNAME -f \$GRIDFILE -t regional
cd ..

```

5.) skip
6.) Generate domain file for datm and CLM
    cd ../../../../tools/mapping/gen_domain_files/src
    gmake
    cd ..
    setenv OCNDOM domain.ocn_nocean.nc
    setenv ATMDOM domain.lnd.{$GRIDNAME}_nocean.nc
    ./gen_domain -m $MAPFILE -o $OCNDOM -l $ATMDOM
    cd ../../../../../lnd/clm/tools
7.) Create surface dataset for CLM
    cd clm4_5/mksurfdata_map/src
    gmake
    cd ..
    ./mksurfdata.pl -r usrspec -usr_gname $GRIDNAME -usr_gdate $CDATE
8.) skip
9.) skip

V. Notes on which input datasets are needed for CLM

global or regional/single-point grids
- need fsurdata and fatmlndfrc

fsurdata ---- from mksurfdata_map in step (III.7)
fatmlndfrc -- use the domain.lnd file from gen_domain in step (III.6)
( NOTE: THIS FILE IS POINTED TO USING ATM_DOMAIN_PATH/ATM_DOMAIN_FILE/LND_DOMAIN_
PATH/ \
LND_DOMAIN_FILE
        env_run.xml variables -- do NOT simply add this to your user_nl_clm as it_
will fail)

```

1.3.5 Datasets for Observational Sites

There are two ways to customize datasets for a particular observational site. The first is to customize the input to the tools that create the dataset, and the second is to over-write the default data after you've created a given dataset. Depending on the tool it might be easier to do it one way or the other. In [Table 3-1](#) we list the files that are most likely to be customized and the way they might be customized. Of those files, the ones you are most likely to customize are: fatmlndfrc, fsurdat, faerdep (for DATM), and stream_fldfilename_ndep. Note **mksurfdata_map** as documented previously has options to overwrite the vegetation and soil types. For more information on this also see [the Section called Creating your own single-point/regional surface datasets in Chapter 5](#). And PTCLM uses these methods to customize datasets see [Chapter 6](#).

Another aspect of customizing your input datasets is customizing the input atmospheric forcing datasets. See [the Section called Running with your own atmosphere forcing in Chapter 5](#) for more information on this. Also the chapter on PTCLM in the Section called Converting AmeriFlux Data for use by PTCLM in [Chapter 6](#) has information on using the AmeriFlux tower site data as atmospheric forcing.

1.3.6 Creating CLM domain files

gen_domain to create a domain file for datm from a mapping file. The domain file is then used by BOTH DATM AND CLM to define the grid and land-mask. The general data flow is shown in two figures. [Figure 2-4](#) shows the general flow for a general global case (or for a regional grid that DOES include ocean). [Figure 2-5](#) shows the use of **mknocnmap.pl** (see [the Section called Using mknocnmap.pl to create grid and maps for single-point regional grids](#)) to create a regional or single-point map file that is then run through **gen_domain** to create the domain file for it. As

stated before Figure 2-2 is the legend for both of these figures. See the [tools/mapping/gen_domain_files/README](#) file for more help on **gen_domain**.

Here we create domain files for a regular global domain.

Figure 2-4. Global Domain file creation

Insert figure 2-4

Starting from SCRIP grid files for both your atmosphere and ocean, you use [tools/mapping/gen_mapping_files/gen_cesm_maps.sh](#) to create a mapping file between the atmosphere and ocean. That mapping file is then used as input to **gen_domain** to create output domain files for both atmosphere and ocean. The atmosphere domain file is then used by both CLM and DATM for I compsets, while the ocean domain file is ignored. For this process you have to define your SCRIP grid files on your own. For a regional or single-point case that doesn't include ocean see [Figure 2-5](#). (See [Figure 2-2](#) for the legend for this figure.)

Note, that the SCRIP grid file used to start this process, is also used in **mkmapdata.sh** (see the Section called Creating mapping files that `mksurfdata_map` will use). Next we create domain files for a single-point or regional domain.

Figure 2-5. Domain file creation using `mknoocnmap.pl`

Insert figure 2-5

For a regular latitude/longitude grid that can be used for regional or single point simulations – you can use **mknoocnmap.pl**. It creates a SCRIP grid file that can then be used as input to **mkmapdata.sh** as well as a SCRIP mapping file that is then input to **gen_domain**. The output of **gen_domain** is a atmosphere domain file used by both CLM and DATM and a ocean domain file that is ignored. (See [Figure 2-2](#) for the legend for this figure.)

In this case the process creates both SCRIP grid files to be used by **mkmapdata.sh** as well as the domain files that will be used by both CLM and DATM.

1.3.7 Comparing History Files

cprnc is a tool shared by both CAM and CLM to compare two NetCDF history files. It differences every field that has a time-axis that is also shared on both files, and reports a summary of the difference. The summary includes the three largest differences, as well as the root mean square (RMS) difference. It also gives some summary information on the field as well. You have to enter at least one file, and up to two files. With one file it gives you summary information on the file, and with two it gives you information on the differences between the two. At the end it will give you a summary of the fields compared and how many fields were different and how many were identical.

Options:

-m = do NOT align time-stamps before comparing

-v = verbose output

-ipr

-jpr

-kpr

See the **cprnc** README file for more details.

Note: To compare files with OUT a time axis you can use the **cprnc.ncl** NCL script in `models/ind/clm/tools/shared/ncl_scripts`. It won't give you the details on the differences but will report if the files are identical or

different.

1.4 Adding New Resolutions

1.4.1 Adding New Resolutions

In the last chapter we gave the details on how to create new files for input into CLM. These files could be either global resolutions, regional-grids or even a single grid point. If you want to easily have these files available for continued use in your development you will then want to include them in the build-namelist database so that build-namelist can easily find them for you. You can deal with them, just by putting the settings in the `user_nl_clm` namelist file, or by using `CLM_USRDAT_NAME`. Another way to deal with them is to enter them into the database for build-namelist, so that build-namelist can find them for you. This keeps one central database for all your files, rather than having multiple locations to keep track of files. If you have a LOT of files to keep track of it also might be easier than keeping track by hand, especially if you have to periodically update your files. If you just have a few quick experiments to try, for a short time period you might be best off using the other methods mentioned above.

There are two parts to adding files to the build-namelist database. The first part is adding new resolution names which is done in the `models/lnd/clm/bld/namelist_files/namelist_definition_clm4_5.xml` file (and in the `models/lnd/clm/bld/config_files/config_definition.xml` file when adding supported single-point datasets). You can then use the new resolution by using `CLM_USRDAT_NAME`. If you also want to be able to give the resolution into `create_newcase` – you'll need to add the grid to the `scripts/ccsm_utils/Case.template/config_grid.xml` file.

The second part is actually adding the new filenames which is done in the `models/lnd/clm/bld/namelist_files/namelist_defaults_clm4_5.xml` file (`models/lnd/clm/bld/namelist_files/namelist_defaults_clm4_5_tools.xml` file for CLM tools). If you aren't adding any new resolutions, and you are just changing the files for existing resolutions, you don't need to edit the `namelist_definition` file.

1.4.2 Managing Your Data Own Files

If you are running on a supported machine (such as yellowstone or hopper) the standard input datasets will already be available and you won't have to check them out of the subversion inputdata server. However, you also will NOT be able to add your own datafiles to these standard inputdata directories – because most likely you won't have permissions to do so. In order to add files to the XML database or to use `CLM_USRDAT_NAME` you need to put data in the standard locations so that they can be found. The recommended way to do this is to use the `link_dirtree` tool in the CESM scripts. Some information on `link_dirtree` is available in the [CESM1.2.0 Scripts User's Guide](#). We also have some examples of it's use here and in other sections of this User's Guide.

Using `link_dirtree` is quite simple, you give the directory where data exists and then the directory that you want to create where datasets will point to the original source files. In the example below we use “`$HOME/inputdata`”, but `MYCSMDATA` could be any directory you have access to where you want to put your data.

```
` > cd scripts # First make sure you have a inputdata location that you can
write to # You only need to do this step once, so you won't need to do this
in the future # (except to bring in any updated files in the original $CSMDATA
location). > setenv MYCSMDATA $HOME/inputdata # Set env var for the directory
for input data > ./link_dirtree $CSMDATA $MYCSMDATA`
```

Then when you create a case you will change `DIN_LOC_ROOT_CSMDATA` to point to the location you linked to rather than the default location.

```
> ./xmlchange DIN_LOC_ROOT_CSMDATA=$MYCSMDATA
```

In order to list the files that you have created you merely need to use the UNIX command find to find the files that are NOT softlinks. So for example executing the following command:

```
> find $MYCSMDATA -type f -print
```

for me gives the following truncated list of CLM_USRDATA_NAME files that I have created.

```
` /glade/p/work/erik/inputdata/lnd/clm2/pftdata/pft-physiology.c130503.nc /
glade/p/work/erik/inputdata/atm/datm7/CLM1PT_data/1x1pt_BE-Vie/1997-01.nc
/glade/p/work/erik/inputdata/atm/datm7/CLM1PT_data/1x1pt_BE-Vie/1997-02.nc
/glade/p/work/erik/inputdata/atm/datm7/CLM1PT_data/1x1pt_BE-Vie/1997-03.nc
/glade/p/work/erik/inputdata/atm/datm7/CLM1PT_data/1x1pt_BE-Vie/1997-04.nc`
```

You can also use **find** to list files that have a particular pattern in the name as well (using the -name option with wildcards). Also you can always rerun the **link_dirtree** command if any new files are added that you need to be linked into your directory tree. Since, the files are soft-links – it doesn't take up much space other than the files that you add there. This way all of the files are kept in one place, they are organized by usage according to CESM standards, and you can easily find your own files, and CLM can find them as well.

If you are running on a supported machine (such as yellowstone or hopper) the standard input datasets will already be available and you won't have to check them out of the subversion inputdata server. However, you also will NOT be able to add your own datafiles to these standard inputdata directories – because most likely you won't have permissions to do so. In order to add files to the XML database or to use CLM_USRDATA_NAME you need to put data in the standard locations so that they can be found. The recommended way to do this is to use the **link_dirtree** tool in the CESM scripts. Some information on **link_dirtree** is available in the [CESM1.2.0 Scripts User's Guide](#). We also have some examples of it's use here and in other sections of this User's Guide.

Using **link_dirtree** is quite simple, you give the directory where data exists and then the directory that you want to create where datasets will point to the original source files. In the example below we use “\$HOME/inputdata”, but MYCSMDATA could be any directory you have access to where you want to put your data.

```
` > cd scripts # First make sure you have a inputdata location that you can
write to # You only need to do this step once, so you won't need to do this
in the future # (except to bring in any updated files in the original $CSMDATA
location). > setenv MYCSMDATA $HOME/inputdata # Set env var for the directory
for input data > ./link_dirtree $CSMDATA $MYCSMDATA`
```

Then when you create a case you will change DIN_LOC_ROOT_CSMDATA to point to the location you linked to rather than the default location.

```
> ./xmlchange DIN_LOC_ROOT_CSMDATA=$MYCSMDATA
```

In order to list the files that you have created you merely need to use the UNIX command find to find the files that are NOT softlinks. So for example executing the following command:

```
> find $MYCSMDATA -type f -print
```

for me gives the following truncated list of CLM_USRDATA_NAME files that I have created.

```
` /glade/p/work/erik/inputdata/lnd/clm2/pftdata/pft-physiology.c130503.nc /
glade/p/work/erik/inputdata/atm/datm7/CLM1PT_data/1x1pt_BE-Vie/1997-01.nc
/glade/p/work/erik/inputdata/atm/datm7/CLM1PT_data/1x1pt_BE-Vie/1997-02.nc
/glade/p/work/erik/inputdata/atm/datm7/CLM1PT_data/1x1pt_BE-Vie/1997-03.nc
/glade/p/work/erik/inputdata/atm/datm7/CLM1PT_data/1x1pt_BE-Vie/1997-04.nc`
```

You can also use **find** to list files that have a particular pattern in the name as well (using the -name option with wildcards). Also you can always rerun the **link_dirtree** command if any new files are added that you need to be linked into your directory tree. Since, the files are soft-links – it doesn't take up much space other than the files that you add there. This way all of the files are kept in one place, they are organized by usage according to CESM standards, and you can easily find your own files, and CLM can find them as well.

1.4.3 Adding Resolution Names

If you are adding files for new resolutions which aren't covered in the namelist_definition file – you'll need to add them in. The list of valid resolutions is in the id="res" entry in the models/lnd/clm/bld/namelist_files/namelist_definition_clm4_5.xml file. You need to choose a name for your new resolution and simply add it to the comma delimited list of valid_values for the id="res" entry. The convention for global Gaussian grids is number_of_latitudes x number_of_longitudes. The convention for global finite volume grids is latitude_grid_size x longitude_grid_size where latitude and longitude is measured in degrees. The convention for unstructured HOMME grids is ne<size>np4, where <size> corresponds to the resolution. The higher <size> is the higher the resolution. So for example, ne60np4 is roughly half-degree while ne240np4 is roughly a eighth degree. For regional or single-point datasets the names have a grid size number_of_latitudes x number_of_longitudes followed by an underscore and then a descriptive name such as a City name followed by an abbreviation for the Country in caps. The only hard requirement is that names be unique for different grid files. Here's what the entry for resolutions looks like in the file:

```
<entry id="res" type="char*30" category="default_settings"
      group="default_settings"
      valid_values=
        "512x1024,360x720cru,128x256,64x128,48x96,32x64,8x16,94x192,0.23x0.31,0.9x1.25,
        ↵1.9x2.5,2.5x3.33,
        4x5,10x15,5x5_amazon,1x1_tropicAtl,1x1_camdenNJ,1x1_vancouverCAN,1x1_
        ↵mexicocityMEX,1x1_asphaltjungleNJ,
        1x1_brazil,1x1_urbanc_alpha,1x1_numaIA,1x1_smallvilleIA,0.1x0.1,0.5x0.5,3x3min,
        ↵5x5min,10x10min,0.33x0.3,
        ne4np4,ne16np4,ne30np4,ne60np4,ne120np4,ne240np4,wus12,us20,1km-merge-10min">
      Horizontal resolutions
      Note: 0.1x0.1, 0.5x0.5, 5x5min, 10x10min, 3x3min, 1km-merge-10min, and 0.33x0.
      ↵33 are only used for CLM tools
</entry>
```

As you can see you just add your new resolution names to the end of the valid_values list.

When using PTCLM and adding supported single-point resolutions, you'll also want to add these resolutions to the models/lnd/clm/bld/config_files/config_definition.xml under the sitespf_pt name. The entry in that file looks like:

```
<entry id="sitespf_pt"
      valid_values="none,1x1_brazil,1x1_tropicAtl,5x5_amazon,1x1_camdenNJ,1x1_
      ↵vancouverCAN,
      1x1_mexicocityMEX,1x1_asphaltjungleNJ,1x1_urbanc_alpha,1x1_numaIA,1x1_
      ↵smallvilleIA,us20,wus12"
      value="none" category="physics">
      Flag to turn on site specific special configuration flags for supported single
      point resolutions. See the specific config_defaults_*.xml file for the special
      settings that are set for a particular site.
</entry>
```

PTCLM assumes that any supported single-point resolutions are valid settings for sitespf_pt.

1.4.4 Changing Default Filenames

To add or change the default filenames you edit the models/lnd/clm/bld/namelist_files/namelist_defaults_clm4_5.xml and either change an existing filename or add a new one. Most entries in the default namelist files, include different attributes that describe the different properties that describe the differences in the datasets. Attributes include the: resolution, year to simulation, range of years to simulate for transient datafiles, the land-mask, the representative concentration pathway (rcp) for future scenarios, and the type of biogeochemistry (bgc) model used. For example the fatmgrid for the 1.9x2.5 resolution is as follows:

```
` <fsurdat hgrid="0.9x1.25" sim_year="1850" crop="off" > lnd/clm2/
surfdatal/surfdatal_0.9x1.25_simyr1850_c130415.nc</fsurdat> `
```

Other fsurdat files are distinguished from this one by their resolution (hgrid), simulation year (sim_year) and prognostic crop (crop) attributes.

To add or change the default filenames for CLM tools edit the models/lnd/clm/bld/namelist_files/namelist_defaults_clm4_5_tools.xml and either change an existing filename or add a new one. Editing this file is similar to the namelist_defaults_clm4_5.xml talked about above.

What are the required files?

Different types of simulations and different types of configurations for CLM require different lists of files. The CLM4.5-BGC or Carbon Nitrogen (cn) Biogeochemistry model for example requires stream_fldfilename_ndep files, which are NOT required by CLMSP. Transient simulations also require transient datasets, and the names of these datasets are sometimes different from the static versions (sometimes both are required as in the dynamic PFT cases).

In the following table we list the different files used by CLM, they are listed in order of importance, dependencies, and customizing. So the required files are all near the top, and the files used only under different conditions are listed later, and files with the fewest dependencies are near the top, as are the files that are least likely to be customized.

Table 3-1. Required Files for Different Configurations and Simulation Types

Insert table 3-1

1.5 Running Special Cases

1.5.1 What is a special case?

All of the following special cases cases that take more than one step to do. The straightforward cases have compsets and/or build-namelist use-cases setup for them or require simple editing of a single-case. All of the cases here require you to do at least two simulations with different configurations, or require more complex editing of the case (changing the streams files).

Note: The cases in this chapter are more sophisticated and require more technical knowledge and skill than cases in previous chapters. The user should be very familiar with doing simple cases before moving onto the cases described here.

1.5.2 Running the prognostic crop model

The prognostic crop model is setup to work with CLM4.5-BGC or CLM4.0-CN (with or without DV) for present day conditions and we have surface and initial condition datasets at f19 resolution. In order to use the initial condition file, we need to set the RUN_TYPE to startup rather than hybrid since the compset for f19 sets up to use an initial condition file without crop active. To activate the crop model you can choose a compset that has “CROP” in the name such as “ICRUCLM45BGCCROP” or simply add “-crop on” to CLM_CONFIG_OPTS.

Example: Crop Simulation

```
> cd scripts
> ./create_newcase -case CROP -res f19_g16 -compset I1850CRUCLM45BGC
> cd CROP

# Append "-crop on" to CLM_CONFIG_OPTS in env_build.xml (you could also use an editor)
> ./xmlchange CLM_CONFIG_OPTS="-crop on" -append

# Change to startup type so uses spunup initial conditions file for crop if it exists
# By default the model will do a hybrid startup with an initial condition file
# incompatible with the crop surface dataset.

> ./xmlchange RUN_TYPE=startup
> ./case.setup

# Now build and run normally
> ./case.build
> ./case.submit
```

1.5.3 Running with irrigation

In CLM4.5 irrigation can ONLY be used WITH crop. To turn on irrigation in CLM4.5 we simply add “-irrig on” to CLM_BLDNML_OPTS. Just as in the crop example we also change RUN_TYPE to startup so that we don’t use an initial condition file that is incompatible with irrigation.

Example: Irrigation Simulation

```
# Note here we do a CLMSP simulation as that is what has been validated
> cd scripts
> ./create_newcase -case IRRIG -res f19_g16 -compset I
> cd IRRIG

# Append "-irrig" to CLM_BLDNML_OPTS in env_run.xml (you could also use an editor)
> ./xmlchange CLM_BLDNML_OPTS="-irrig" -append

# Change to startup type so uses spunup initial conditions file for irrigation if it_
# exists
# By default the model will do a hybrid startup with an initial condition file
# incompatible with the irrigation surface dataset.
> ./xmlchange RUN_TYPE=startup
> ./case.setup

# Now build and run normally
> ./case.build
> ./case.submit
```

1.5.4 Spinning up the Satellite Phenology Model

To spin-up the CLMSP model you merely need to run CLMSP for 50 simulation years starting from arbitrary initial conditions. You then use the final restart file for initial conditions in other simulations. Because, this is a straight forward operation we will NOT give the details on how to do that here, but leave it as an exercise for the reader. See the [Example 4-7](#) as an example of doing this as the last step for CLMCN.

1.5.5 Spinup of CLM4.5-BGC

To get the CLM4.5-BGC model to a steady state, you first run it from arbitrary initial conditions using the “accelerated decomposition spinup” (-bgc_spinup on in CLM **configure**) mode for 1000 simulation years. After this you branch from this mode in the “final spinup” (-bgc_spinup off in CLM **configure**), and run for (at least 200+ simulation years).

1. 45_AD_SPINUP For the first step of running 1000+ years in “-bgc_spinup on” mode, you will setup a case, and then edit the values in env_build.xml and env_run.xml so that the right configuration is turned on and the simulation is setup to run for the required length of simulation time. So do the following:

Example:: AD_SPINUP Simulation for CLM4.5-BGC

```
> cd scripts
> ./create_newcase -case BGC_spinup -res f19_g16 -compset I1850CRUCLM45BGC -mach_
↪yellowstone_intel
> cd BGC_spinup
# Append "-spinup on" to CLM_BLDNML_OPTS
> ./xmlchange CLM_BLDNML_OPTS="-bgc_spinup on" -append
# The following sets CLM_FORCE_COLDSTART to "on", and run-type to startup (you could_
↪also use an editor)
> ./xmlchange CLM_FORCE_COLDSTART=on,RUN_TYPE=startup
# Make the output history files only annual, by adding the following to the user_nl_
↪clm namelist
> echo 'hist_nhtfrq = -8760' >> user_nl_clm
# Now setup
> ./cesm_setup -case
# Now build
> ./BGC_spinup.build
# The following sets RESUBMIT to 30 times in env_run.xml (you could also use an_
↪editor)
# The following sets STOP_DATE,STOP_N and STOP_OPTION to Jan/1/1001, 20, "nyears" in_
↪env_run.xml (you could also use an editor)
> ./xmlchange RESUBMIT=20,STOP_N=50,STOP_OPTION=nyears,STOP_DATE=10010101
# Now run normally
> ./BGC_spinup.submit
```

Note: This same procedure works for CLM4.5-CN as well, you can typically shorten the spinup time from 1000 years to 600 though.

Afterwards save the last restart file from this simulation to use in the next step.

2. Final spinup for CLM4.5-BGC Next save the last restart file from this step and use it as the “finidat” file to use for one more spinup for at least 200+ years in normal mode. So do the following:

Example: Final CLMBGC Spinup Simulation for CLM4.5-BGC

```
> cd scripts
> ./create_newcase -case BGC_finalspinup -res f19_g16 -compset I1850CRUCLM45BGC -mach_
↪yellowstone_intel
> cd BGC_finalspinup
# Now, Copy the last CLM restart file from the earlier case into your run directory
> cp /ptmp/$LOGIN/archive/BGC_spinup/rest/BGC_spinup.clm*.r*.1002-01-01-00000.nc \
/glade/scratch/$LOGIN/CN_finalspinup
# Set the runtype to startup
```

```

> ./xmlchange RUN_TYPE=startup
# And copy the rpointer files for datm and drv from the earlier case
> cp /glade/scratch/$LOGIN/archive/BGC_spinup/rest/rpointer.atm /ptmp/$LOGIN/CN_
˓→finalspinup
> cp /glade/scratch/$LOGIN/archive/BGC_spinup/rest/rpointer.drv /ptmp/$LOGIN/CN_
˓→finalspinup
# Set the finidat file to the last restart file saved in previous step
> echo ' finidat = "BGC_spinup.clm2.r.1002-01-01-00000.nc"' > user_nl_clm
# Now setup
> ./cesm_setup
> Now build
> ./BGC_finalspinup.build
# The following sets RESUBMIT to 4 times in env_run.xml (you could also use an editor)
# The following sets STOP_N and STOP_OPTION to 50 and "nyears" in env_run.xml (you
˓→could also use an editor)
> ./xmlchange RESUBMIT=4,STOP_OPTION=nyears,STOP_N=50
> Now run as normal
> ./BGC_finalspinup.submit

```

To assess if the model is spunup plot trends of CLMBGC variables of interest. If you see a trend, you may need to run the simulation longer. Finally save the restart file from the end of this simulation to use as an “finidat” file for future simulations.

Note: This same final spinup procedure works for CLM4.5-CN as well, you can typically shorten the spinup time from 200 years to 50 though.

1.5.6 Running with MOAR data

Because it takes so long to spinup the CN model (as we just saw previously), if you are doing fully coupled simulations with active atmosphere and ocean, you will want to do the spinup portion of this “offline”. So instead of doing expensive fully coupled simulations for the spinup duration, you run CLM in a very cheap “I” compset using atmospheric forcing from a shorter fully coupled simulation (or a simulation run previously by someone else).

In this example we will use the I1850SPINUPCN compset to setup CLM to run with atmospheric forcing from a previous fully coupled simulation with data that is already stored on disk on yellowstone. There are several simulations that have high frequency data for which we can do this. You can also do this on a machine other than yellowstone, but would need to download the data from the Earth System Grid and change the datapath similar to [Example 4-11](#).

Example: Simulation with MOAR Data on yellowstone

```

> cd scripts
> ./create_newcase -case MOARforce1850 -res f19_g16 -compset I1850SPINUPCN -mach_
˓→yellowstone_intel
> cd MOARforce1850
# The following sets the casename to point to for atm forcing (you could also use an
˓→editor)
> ./xmlchange DATM_CPL_CASE=b40.1850.track1.1deg.006a
# The following sets the align year and years to run over for atm forcing
# (you could also use an editor)
> ./xmlchange DATM_CPL_YR_ALIGN=1,DATM_CPL_YR_START=960,DATM_CPL_YR_END=1030
> ./cesm_setup
# Now build and run as normal

```

```
> ./MOARforce1850.build  
> ./MOARforce1850.submit
```

1.5.7 Running with atmospheric forcing from a previous simulation

Another way that you might want to spinup the model is to run your own simulation for a relatively short period (either a B, E, or F compset) and then use it as forcing for your “I” case later. By only running 20 to 50 years for the fully coupled case, you’ll save a substantial amount of computer time rather than running the entire spinup period with a fully coupled model.

The first thing we need to do is to run a fully coupled case and save the atmospheric coupling fields on a three hourly basis. In this example, we will run on yellowstone and archive the data to a local disk that we can then use in the next simulation.

Example: Fully Coupled Simulation to Create Data to Force Next Example Simulation

```
> cd scripts  
> ./create_newcase -case myBCN1850 -res f09_g16 -compset B1850CN -mach yellowstone_  
→intel  
> cd myBCN1850  
> ./cesm_setup  
# Set histaux_a2x3hr to .true. in your user_nl_cpl output from the atmosphere model  
# will be saved 3 hourly  
echo "histaux_a2x3hr=.true." >> user_nl_cpl  
# edit the driver code in order to save the correct list of fields (see note below)  
> cp ../../models/driv/driver/ccsm_comp_mod.F90 SourceMods/src.cpl  
> $EDITOR SourceMods/src.cpl  
# Now build  
> ./myBCN1850.build  
# The following sets the archival disk space (you could also use an editor)  
> ./xmlchange DOUT_S_ROOT='/glade/home/$USER/$CASE'  
# Make sure files are archived to disk, but NOT to long term storage  
# (you could also use an editor)  
> ./xmlchange DOUT_S=TRUE,DOUT_L_MS=FALSE  
# Set the run length to run a total of 20 years (you could also use an editor)  
> ./xmlchange RESUBMIT=9,STOP_OPTION=nyears,STOP_N=2  
# Now run as normal  
> ./myBCN1850.submit
```

Warning: Because of bug 1733 (see the [models/lnd/clm/doc/KnownBugs](#) file on this) you’ll need to edit the driver code in order for it to produce the correct list of fields needed to run the model later.

Now we run an I compset forced with the data from the previous simulation using the CPLHIST3HrWx option to DATM_MODE. See the [Section called CPLHIST3HrWx mode and its DATM settings in Chapter 1](#) for more information on the DATM settings for CPLHIST3HrWx mode.

Example: Simulation Forced with Data from the Previous Simulation

```
> cd scripts  
> ./create_newcase -case frcwmyBCN1850 -res f09_g16 -compset I1850SPINUPCN -mach_  
→yellowstone_intel
```

```
> cd frcWmyBCN1850
# The following sets the casename to point to for atm forcing (you could also use an
# editor)
> ./xmlchange DATM_CPLHIST_CASE="myBCN1850"
# The following sets the align year and years to run over for atm forcing
# (you could also use an editor)
> ./xmlchange DATM_CPLHIST_YR_ALIGN="1",DATM_CPLHIST_YR_START=1,DATM_CPLHIST_YR_END=20
# Set the strm_datadir in the namelist_defaults_datm.xml
# file to the archival path of the case above in the form of: /glade/home/archive/
# $USER/$DATM_CPLHIST_CASE/cpl/hist
# NOTE: THIS WILL CHANGE THE PATH FOR ALL I1850SPINUPCN COMPSET CASES MADE AFTER THIS!
> $EDITOR ../../models/atm/datm/bld/namelist_files/namelist_defaults_datm.xml
> ./cesm_setup
# Now build and run as normal
> ./frcwmyBCN1850.build
> ./frcwmyBCN1850.submit
```

Note: We did this by editing the “namelist_defaults_datm.xml” which will change the settings for ALL future I1850SPINUPCN cases you run. You could also do this by editing the path in the resulting streams text files in the CaseDocs directory, and then create a “user_” streams file with the correct path. This would change the streams file JUST for this case. The steps do it this way are:

```
> ./preview_namelists
> cp CaseDocs/datm.streams.txt.CPLHIST3HrWx.Precip           user_datm.streams.txt.
# CPLHIST3HrWx.Precip
> cp CaseDocs/datm.streams.txt.CPLHIST3HrWx.Solar            user_datm.streams.txt.
# CPLHIST3HrWx.Solar
> cp CaseDocs/datm.streams.txt.CPLHIST3HrWx.nonSolarNonPrecip user_datm.streams.txt.
# CPLHIST3HrWx.nonSolarNonPrecip
# Change the <fieldInfo> field <filePath> to point to the correct directory i.e.: /
# glade/home/archive/$USER/$DATM_CPLHIST_CASE/cpl/hist
> $EDITOR user_datm.streams.txt.CPLHIST3HrWx./*
> ./preview_namelists
# Then make sure the CaseDocs/datm.streams.txt.CPLHIST3HrWx.* files have the correct
# path
```

1.5.8 Running with historical CO2 forcing

In this case you want to run a simulation with stand-alone CLM responding to changes in CO2 for a historical period. For this example, we will start with the “I_1850-2000_CN” compset that has transient: land-use, Nitrogen and Aerosol deposition already. You could also use another compset if you didn’t want these other features to be transient. In order to get CO2 to be transient we need to add a new streams file and add it to the list of streams in the user_nl_datm file. You also need a NetCDF datafile that datm can read that gives the variation. You could supply your own file, but we have a standard file that is used by CAM for this and our example will make use of this file.

Note: Most everything here has to do with changing datm rather than CLM to allow this to happen. As such the user that wishes to do this should first become more familiar with datm and read the [CESM Data Model User’s Guide](#) especially as it pertains to the datm.

Warning: This section documents the process for doing something that is non-standard. There may be errors with the documentation and process, and you may have to do some work before all of this works for you. If that is the case, we recommend that you do further research into understanding the process and the files, as well as understanding the datm and how it works. You may have to read documentation found in the code for datm as well as “csm_share”.

The datm has “streams” files that have rough XML-like syntax and specify the location and file to get data from, as well as information on the variable names and the data locations of the grid points. The datm expects specific variable names and the datm “maps” the expected variable names from the file to the names expected by datm. The file we are working with here is a file with a single-point, that covers the entire globe (so the vertices go from -90 to 90 degrees in latitude and 0 to 360 degrees in longitude). Since it’s a single point it’s a little easier to work with than datasets that may be at a given horizontal resolution. The datm also expects that variables will be in certain units, and only expects a limited number of variables so arbitrary fields can NOT be exchanged this way. However, the process would be similar for datasets that do contain more than one point.

The three things that are needed: a domain file, a data file, and a streams text file. The domain file is a CF-compliant NetCDF file that has information on the grid points (latitudes and longitudes for cell-centers and vertices, mask, fraction, and areas). The datafile is a CF-compliant NetCDF file with the data that will be mapped. The streams text file is the XML-like file that tells datm how to find the files and how to map the variables datm knows about to the variable names on the NetCDF files. Note, that in our case the domain file and the data file are the same file. In other cases, the domain file may be separate from the data file.

First we are going to create a case, and we will edit the user_nl_datm so that we add a CO2 data stream in. There is a streams text file available in models/lnd/clm/doc/UsersGuide/co2_streams.txt, that includes file with a CO2 time-series from 1765 to 2007.

Example: Transient Simulation with Historical CO2

```
> cd scripts
> ./create_newcase -case DATM_CO2_TSERIES -res f19_g16 -compset I20TRCRUCLM45BGC
> cd DATM_CO2_TSERIES

# Set CCSM_BGC to CO2A so that CO2 will be passed from atmosphere to land
# Set CLM_CO2_TYPE to diagnostic so that the land will use the value sent from the
# atmosphere
> ./xmlchange CCSM_BGC=CO2A,CLM_CO2_TYPE=diagnostic
> ./case.setup

# Create the streams file for CO2
> cat << EOF >> datm.streams.txt.co2tseries

<streamstemplate>
<general_comment>
  This is a streams file to pass historical CO2 from datm8 to the other
  surface models. It reads in a historical dataset derived from data used
  by CAM. The getco2_historical.ncl script in models/lnd/clm2/tools/ncl_scripts
  was used to convert the CAM file to a streams compatible format (adding domain
  information and making CO2 have latitude/longitude even if only for a single
  point.
</general_comment>
<stream>
<comment>
  Input stream description file for historical CO2 reconstruction data
  04 March 2010: Converted to form that can be used by datm8 by Erik Kluzek

```

```

18 December 2009: Prepared by B. Eaton using data provided by
Jean-Francois Lamarque. All variables except f11 are directly from
PRE2005_MIDYR_CONC.DAT. Data from 1765 to 2007 with 2006/2007 just
a repeat of 2005.
</comment>
<dataSource>
  CLMNCEP
</dataSource>
<domainInfo>
  <variableNames>
    time   time
    lonc   lon
    latc   lat
    area   area
    mask   mask
  </variableNames>
  <filePath>
    $CSMDATA/atm/datm7/CO2
  </filePath>
  <fileNames>
    fco2_datm_1765-2007_c100614.nc
  </fileNames>
</domainInfo>
<fieldInfo>
  <variableNames>
    CO2      co2diag
  </variableNames>
  <filePath>
    $CSMDATA/atm/datm7/CO2
  </filePath>
  <fileNames>
    fco2_datm_1765-2007_c100614.nc
  </fileNames>
</fieldInfo>
</stream>
</streamstemplate>

EOF

# And copy it to the run directory
> cp datm.streams.txt.co2tseries $RUNDIR

# Run preview namelist so we have the namelist in CaseDocs
> ./preview_namelists

```

The first thing we will do is to edit the `user_nl_datm` file to add a CO2 file stream in. To do this we will copy a `user_nl_datm` in with the changes needed. The file `addco2_user_nl_datm.user_nl` is in `models/lnd/clm/doc/UsersGuide` and looks like this...

```

dtlimit = 1.5,1.5,1.5,1.5,1.5
fillalgo = 'nn','nn','nn','nn','nn'
fillmask = 'nomask','nomask','nomask','nomask','nomask'
mapalgo = 'bilinear','bilinear','bilinear','bilinear','nn'
mapmask = 'nomask','nomask','nomask','nomask','nomask'
streams = "datm.streams.txt.CLM_QIAN.Solar 1895 1948 1972 ", "datm.streams.txt.CLM_
↪QIAN.Precip 1895 1948 1972 ",
          "datm.streams.txt.CLM_QIAN.TPQW 1895 1948 1972 ", "datm.streams.txt.
↪presaero.trans_1850-2000 1849 1849 2006",

```

```
"datm.streams.txt.co2tseries 1766 1766 2005 "
taxmode = 'cycle','cycle','cycle','cycle','extend'
tintalgo = 'coszen','nearest','linear','linear','linear'
```

You just copy this into your case directory. But, also compare it to the version in CaseDocs to make sure the changes are just to add in the new CO2 stream. Check to see that filenames, and start, end and align years are correct.

```
> cp ../../models/lnd/clm/doc/UsersGuide/addco2_user_nl_datm.user_nl user_nl_datm
> diff user_nl_datm CaseDocs/datm_atm_in
```

Once, you've done that you can build and run your case normally.

Warning: This procedure assumes you are using a I20TRCRUCLM45BGC compset out of the box, with DATM_PRESAERO equal to trans_1850-2000. So it assumes standard CLM4.5 CRUNCEP atmosphere forcing, and transient prescribed aerosols from streams files. If your case changes anything here your user_nl_datm file will need to be adjusted to work with it.

Note: The intent of the user_nl_datm is to add an extra streams file for CO2 to the end of the streams variable, and other arrays associated with streams (adding mapalgo as a new array with bilinear for everything, but the CO2 file which should be “nn” for nearest neighbor). Other variables should be the same as the other stream values.

Warning: The streams file above is hard-coded for the path of the file on NCAR computers. To use it on an outside machine you'll need to edit the filepath in the streams file to point to the location where you have the file.

After going through these steps, you will have a case where you have datm reading in an extra streams text file that points to a data file with CO2 data on it that will send that data to the CLM.

1.6 Running Single Point Regional Cases

1.6.1 Single and Regional Grid Configurations

CLM allows you to set up and run cases with a single-point or a local region as well as global resolutions. This is often useful for running quick cases for testing, evaluating specific vegetation types, or land-units, or running with observed data for a specific site.

There are three different ways to do this for normal-supported site

PTS_MODE runs for a single point using global datasets.

CLM_USRDAT_NAME runs using your own datasets (single-point or regional).

PTCLM easily setup simulations to run for tower sites..

Note: PTS_MODE and PTCLM only works for a single point, while the other two options can also work for regional datasets as well.

Choosing the right single point options

Running for a *normal supported site* is a great solution, if one of the supported single-point/regional datasets, is your region of interest (see [the Section called Running Supported Single-point/Regional Datasets](#)). All the datasets are created for you, and you can easily select one and run, out of the box with it using a supported resolution from the top level of the CESM scripts. The problem is that there is a very limited set of supported datasets. You can also use this method for your own datasets, but you have to create the datasets, and add them to the XML database in scripts, CLM and to the DATM. This is worthwhile if you want to repeat many multiple cases for a given point or region.

In general [the Section called Running PTS_MODE configurations](#) is the quick and dirty method that gets you started without having to create datasets – but has limitations. It's good for an initial attempt at seeing results for a point of interest, but since you can NOT restart with it, it's usage is limited. It is the quickest method as you can create a case for it directly from **create_newcase**. Although you can't restart, running a single point is very fast, and you can run for long simulation times even without restarts.

Next, `CLM_USRDAT_NAME` is the best way to setup cases quickly where you have to create your own datasets (see [the Section called Creating your own single-point/regional surface datasets](#)). With this method you don't have to change DATM or add files to the XML database – but you have to follow a strict naming convention for files. However, once the files are named and in the proper location, you can easily setup new cases that use these datasets. This is good for treating all the required datasets as a “group” and for a particular model version. For advanced CLM developers who need to track dataset changes with different model versions you would be best off adding these datasets as supported datasets with the “normal supported datasets” method.

Lastly *PTCLM* is a great way to easily create datasets, setup simulations and run simulations for tower sites. It takes advantage of both normal supported site functionality and `CLM_USRDAT_NAME` internally. A big advantage to it, is that it's one-stop shopping, it runs tools to create datasets, and runs **create_newcase** and sets the appropriate env variables for you. So you only have to learn how to run one tool, rather than work with many different ones. PTCLM is described in the next chapter [Chapter 6](#).

Finally, if you also have meteorology data that you want to force your CLM simulations with you'll need to setup cases as described in [the Section called Running with your own atmosphere forcing](#). You'll need to create CLM datasets either according to `CLM_USRDAT_NAME`. You may also need to modify DATM to use your forcing data. And you'll need to change your forcing data to be in a format that DATM can use. In the PTCLM chapter [the Section called Converting AmeriFlux Data for use by PTCLM in Chapter 6](#) section tells you how to use AmeriFlux data for atmospheric forcing.

1.6.2 Running a single point using global data - PTS_MODE

`PTS_MODE` enables you to run the model using global datasets, but just picking a single point from those datasets and operating on it. It can be a very quick way to do fast simulations and get a quick turnaround.

To setup a `PTS_MODE` simulation you use the “`-pts_lat`” and “`-pts_lon`” arguments to **create_newcase** to give the latitude and longitude of the point you want to simulate for (the code will pick the point on the global grid nearest to the point you give. Here's an example to setup a simulation for the nearest point at 2-degree resolution to Boulder Colorado.

```
> cd scripts
> ./create_newcase -case testPTS_MODE -res f19_g16 -compset I1850CRUCLM45BGC -pts_lat_
→40.0 -pts_lon -105
> cd testPTS_MODE

# We make sure the model will start up cold rather than using initial conditions
> ./xmlchange CLM_FORCE_COLDSTART=on, RUN_TYPE=startup
```

Then setup, build and run as normal. We make sure initial conditions are NOT used since `PTS_MODE` currently CAN NOT run with initial conditions.

Note: By default it sets up to run with MPILIB=mpi-serial (in the env_build.xml file) turned on, which allows you to run the model interactively. On some machines this mode is NOT supported and you may need to change it to FALSE before you are able to build.

Warning: PTS_MODE currently does NOT restart nor is it able to startup from global initial condition files. See bugs “1017 and 1025” in the [models/lnd/clm/doc/KnownLimitations](#) file.

Note: You can change the point you are simulating for at run-time by changing the values of PTS_LAT and PTS_LON in the `env_run.xml` file.

Running on in a single processor

Note, that when running with PTS_MODE the number of processors is automatically set to one. When running a single grid point you can only use a single processor. You might also want to set the `env_build.xml` variable: `MPILIB=mpi-serial` to TRUE so that you can also run interactively without having to use MPI to start up your job.

On many machines, batch queues have a minimum number of nodes or processors that can be used. On these machines you may have to change the queue and possibly the time-limits of the job, to get it to run in the batch queue. On the NCAR machine, yellowstone, this is done for you automatically, and the “caldera” queue is used for such single-processor simulations. Another way to get around this problem is to run the job interactively using `MPILIB=mpi-serial` so that you don’t submit the job to the batch queue. For single point mode you also may want to consider using a smaller workstation or cluster, rather than a super-computer, because you can’t take advantage of the multi-processing power of the super-computer anyway.

1.6.3 Running Single Point Configurations

In addition to PTS_MODE, CLM supports running using single-point or regional datasets that are customized to a particular region. CLM supports a small number of out-of-the-box single-point and regional datasets. However, users can create their own dataset.

To get the list of supported dataset resolutions do this:

```
> cd models/lnd/clm/doc  
> ../../bld/build-namelist -res list
```

Which results in the following:

```
CLM build-namelist - valid values for res (Horizontal resolutions  
Note: 0.1x0.1, 0.5x0.5, 5x5min, 10x10min, 3x3min and 0.33x0.33 are only used for CLM_  
→tools):  
      Values: default 512x1024 360x720cru 128x256 64x128 48x96 32x64 8x16 94x192 \\\n          0.23x0.31 0.47x0.63 0.9x1.25 1.9x2.5 2.5x3.33 4x5 10x15 5x5_amazon 1x1_\\  
→tropicAtl 1x1_camdenNJ \\\n          1x1_vancouverCAN 1x1_mexicocityMEX 1x1_asphaltjungleNJ 1x1_brazil 1x1_\\  
→urbanc_alpha 1x1_numaIA \\\n          1x1_smallvilleIA 0.1x0.1 0.5x0.5 3x3min 5x5min 10x10min 0.33x0.33_\\  
→ne4np4 ne16np4 ne30np4 ne60np4 \\\n          ne120np4 ne240np4 wus12 us20
```

```
Default = 1.9x2.5
```

(NOTE: resolution **and** mask **and** other settings may influence what the default **is**)

The resolution names that have an underscore in them (“_”) are all single-point or regional resolutions.

To run for the Brazil test site do the following:

Example: Running CLM over a single-point test site in Brazil with the default Qian atmosphere data forcing.

```
> cd scripts
> set SITE=1x1_brazil
> ./create_newcase -case testSPDATASET -res $SITE -compset I
> cd testSPDATASET
```

Then setup, build and run normally.

Then to run for the urban Mexico City Mexico test site that also has atmosphere forcing data, but to run it with the Qian forcing data, but over the period for which it's own forcing data is provided do the following:

Example: Running CLM over the single-point of Mexicocity Mexico with the default Qian atmosphere data forcing.

```
> cd scripts
# Set a variable to the site you want to use (as it's used several times below)
> set SITE=1x1_mexicocityMEX
> ./create_newcase -case testSPDATASET -res $SITE -compset I
> cd testSPDATASET
# Set DATM prescribed aerosols to single-point dataset
# Will then use the dataset with just the point for this $SITE
> ./xmlchange DATM_PRESAERO=pt1_pt1
```

Then setup, build and run normally.

Important: Just like PTS_MODE above, By default it sets up to run with MPILIB=mpi-serial (in the env_build.xml file) turned on, which allows you to run the model interactively. On some machines this mode is NOT supported and you may need to change it to FALSE before you are able to build.

Warning: See the Section called Warning about Running with a Single-Processor on a Batch Machine for a warning about running single-point jobs on batch machines.

Note: When running a pt1_pt1 resolution the number of processors is automatically set to one. When running a single grid point you can only use a single processor. You might also want to set the env_build.xml variable: MPILIB=mpi-serial to TRUE so that you can also run interactively without having to use mpi to start up your job.

Using Supported Single-point Datasets that have their own Atmospheric Forcing

Of the supported single-point datasets we have three that also have atmospheric forcing data that go with them: Mexico City (Mexico), Vancouver, (Canada, British Columbia), and urbanc_alpha (test data for an Urban inter-comparison

project). Mexico city and Vancouver also have “#ifdef” in the source code for them to work with modified urban data parameters that are particular to these locations. They can be turned on by using the `CLM_CONFIG_OPTS` env_build.xml variable to set the “-sitespf_pt” option in the CLM **configure**. To turn on the atmospheric forcing for these datasets, you set the `env_run.xml DATM_MODE` variable to `CLM1PT`, and then the atmospheric forcing datasets will be used for the point picked.

When running with datasets that have their own atmospheric forcing you need to be careful to run over the period that data is available. If you have at least one year of forcing it will cycle over the available data over and over again no matter how long of a simulation you run. However, if you have less than a years worth of data (or if the start date doesn’t start at the beginning of the year, or the end date doesn’t end at the end of the year) then you won’t be able to run over anything but the data extent. In this case you will need to carefully set the `RUN_STARTDATE`, `START_TOD` and `STOP_N/STOP_OPTION` variables for your case to run over the entire time extent of your data. For the supported data points, these values are in the XML database and you can use the `queryDefaultNamelist.pl` script to query the values and set them for your case (they are set for the three urban test cases: Mexicocity, Vancouver, and urbanc_alpha).

In the example below we will show how to do this for the Vancouver, Canada point.

Example: Running CLM over the single-point of Vancouver Canada with supplied atmospheric forcing data for Vancouver.

```
> cd scripts

# Set a variable to the site you want to use (as it's used several times below)
> set SITE=1x1_vancouverCAN

# Create a case at the single-point resolutions with their forcing
> ./create_newcase -case testSPDATASETnAtmForcing -res $SITE -compset I1PTCLM45
> cd testSPDATASETnAtmForcing

# Set namelist options for urban test site
> ./xmlchange CLM_NML_USE_CASE=stdurbpt_pd

# Figure out the start and end date for this dataset
# You can do this by examining the datafile.
> set STOP_N=330
> set START_YEAR=1992
> set STARTDATE=${START_YEAR}-08-12
> @ NDAYS = $STOP_N / 24
> ./xmlchange RUN_STARTDATE=$STARTDATE,STOP_N=$STOP_N,STOP_OPTION=nsteps

# Set the User namelist to set the output frequencies of the history files
# Setting the stdurbpt use-case option create three history file streams
# The frequencies and number of time-samples needs to be set
> cat << EOF > user_nl_clm
hist_mfilt = $NDAYS,$STOP_N,$STOP_N
hist_nhtfrq = -1,1,1
EOF

# Set DATM prescribed aerosols to single-point dataset
# Will then use the dataset with just the point for this site
> ./xmlchange DATM_PRESAERO=pt1_pt1
> ./case.setup
```

Warning: If you don't set the start-year and run-length carefully as shown above the model will abort with a "dtlimit error" in the atmosphere model (see bug 1110 in the [models/lnd/clm/doc/KnownLimitations](#) file for documentation on this). Since, the forcing data for this site (and the MexicoCity site) is less than a year, the model won't be able to run for a full year. The `1x1_urbanc_alpha` site has data for more than a full year, but neither year is complete hence, it has the same problem (see the problem for this site above).

Note: Just like `PTS_MODE` above, By default it sets up to run with `MPILIB=mpi-serial` (in the `env_build.xml` file) turned on, which allows you to run the model interactively.

Note: When running a `pt1_pt1` resolution the number of processors is automatically set to one. When running a single grid point you can only use a single processor. You might also want to set the `env_build.xml` variable: `MPILIB=mpi-serial` to TRUE so that you can also run interactively without having to use mpi to start up your job.

Creating your own single-point dataset

The following provides an example of setting up a case using `CLM_USRDAT_NAME` where you rename the files according to the `CLM_USRDAT_NAME` convention. We have an example of such datafiles in the repository for a specific region over Alaska (actually just a sub-set of the global f19 grid).

Example: Using `CLM_USRDAT_NAME` to run a simulation using user datasets for a specific region over Alaska

```
> cd scripts
> ./create_newcase -case my_userdataset_test -res CLM_USRDAT -compset ICRUCLM45
> cd my_userdataset_test/
> set GRIDNAME=13x12pt_f19_alaskaUSA
> set LMASK=gx1v6
> ./xmlchange CLM_USRDAT_NAME=$GRIDNAME,CLM_BLDNML_OPTS="--mask $LMASK"
> ./xmlchange ATM_DOMAIN_FILE=domain.lnd.${GRIDNAME}__$LMASK.nc
> ./xmlchange LND_DOMAIN_FILE=domain.lnd.${GRIDNAME}__$LMASK.nc

# Make sure the file exists in your $CSMDATA or else use svn to download it there
> ls $CSMDATA/lnd/clm2/surfdata_map/surfdata_${GRIDNAME}_simyr2000.nc

# If it doesn't exist, comment out the following...
#> setenv SVN_INP_URL https://svn-ccsm-inputdata.cgd.ucar.edu/trunk/inputdata/
#> svn export $SVN_INP_URL/lnd/clm2/surfdata_map/surfdata_${GRIDNAME}_simyr2000.nc
#> $CSMDATA/lnd/clm2/surfdata_map/surfdata_${GRIDNAME}_simyr2000.nc
> ./case.setup
```

The first step is to create the domain and surface datasets using the process outlined in the Section called The File Creation Process in Chapter 2. Below we show an example of the process.

Example: Creating a surface dataset for a single point

```
# set the GRIDNAME and creation date that will be used later
> setenv GRIDNAME 1x1_boulderCO
> setenv CDATE `date +%y%m%d`
# Create the SCRIP grid file for the location and create a unity mapping file for it.
> cd models/lnd/clm/tools/shared/mkmapdata
> ./mknoocnmap.pl -p 40,255 -n $GRIDNAME
# Set pointer to MAPFILE just created that will be used later
> setenv MAPFILE `pwd`/map_${GRIDNAME}_nocean_to_${GRIDNAME}_nomask_aave_da_${CDATE}.
->nc
# create the mapping files needed by mksurfdata_map.
> cd ../../shared/mkmapdata
> setenv GRIDFILE ../../mkmapgrids/SCRIPgrid_${GRIDNAME}_nomask_${CDATE}.nc
> ./mkmapdata.sh -r $GRIDNAME -f $GRIDFILE -t regional
# create the domain file
> cd ../../../../tools/mapping/gen_domain_files/src
> ../../../../../scripts/ccsm_utils/Machines/configure -mach yellowstone -compiler intel
> gmake
> cd ..
> setenv OCNDOM domain.ocn_nocean.nc
> setenv ATMDOM domain.lnd.${GRIDNAME}_nocean.nc
> ./gen_domain -m $MAPFILE -o $OCNDOM -l $ATMDOM
# Save the location where the domain file was created
> setenv GENDOM_PATH `pwd`
# Finally create the surface dataset
> cd ../../../../lnd/clm/tools/clm4_5/mksurfdata_map/src
> gmake
> cd ..
> ./mksurfdata.pl -r usrspec -usr_gname $GRIDNAME -usr_gdate $CDATE
```

The next step is to create a case that points to the files you created above. We will still use the `CLM_USRDATA_NAME` option as a way to get a case setup without having to add the grid to scripts.

Example: Setting up a case from the single-point surface dataset just created

```
# First setup an environment variable that points to the top of the CESM directory.
> setenv CESMROOT <directory-of-path-to-main-cesm-directory>
# Next make sure you have a inputdata location that you can write to
# You only need to do this step once, so you won't need to do this in the future
> setenv MYCSMDATA $HOME/inputdata      # Set env var for the directory for input data
> ./link_dirtree $CSMDATA $MYCSMDATA
# Copy the file you created above to your new $MYCSMDATA location following the
->CLMUSRDATA
# naming convention (leave off the creation date)
> cp $CESMROOT/models/lnd/clm/tools/clm4_5/mksurfdata_map/surfdata_${GRIDNAME}_
->simyr1850_${CDATE}.nc \
$MYCSMDATA/lnd/clm2/surfdata_map/surfdata_${GRIDNAME}_simyr1850.nc
> cd $CESMROOT/scripts
> ./create_newcase -case my_usernldatasets_test -res CLM_USRDATA -compset_
->I1850CRUCLM45BGC \
-mach yellowstone_intel
> cd my_usernldatasets_test
> ./xmlchange DIN_LOC_ROOT=$MYCSMDATA
# Set the path to the location of gen_domain set in the creation step above
> ./xmlchange ATM_DOMAIN_PATH=$GENDOM_PATH,LND_DOMAIN_PATH=$GENDOM_PATH
```

```
> ./xmlchange ATM_DOMAIN_FILE=$ATMDOM, LND_DOMAIN_FILE=$ATMDOM
> ./xmlchange CLM_USRDATA_NAME=$GRIDNAME
> ./case.setup
```

Note: With this and previous versions of the model we recommended using `CLM_USRDATA_NAME` as a way to identify your own datasets without having to enter them into the XML database. This has two down-sides. First you can't include creation dates in your filenames, which means you can't keep track of different versions by date. It also means you HAVE to rename the files after you created them with `mksurfdatal.pl`. And secondly, you have to use `linkdirtree` in order to place the files in a location outside of the usual `DIN_LOC_ROOT` (assuming you don't have write access to adding new files to the standard location on the machine you are using). Now, since `user_nl` files are supported for ALL model components, and the same domain files are read by both CLM and DATM and set using the envxml variables: `ATM_DOMAIN_PATH`, `ATM_DOMAIN_FILE`, `LND_DOMAIN_PATH`, and `LND_DOMAIN_FILE` – you can use this mechanism (`user_nl_clm` and `user_nl_datm` and those envxml variables) to point to your datasets in any location. In the future we will deprecate `CLM_USRDATA_NAME` and recommend `user_nl_clm` and `user_nl_datm` and the `DOMAIN` envxml variables.

1.7 Running PTCLM

1.7.1 What is PTCLM

PTCLM (pronounced either as point clime or Pee-Tee clime) is a Python script to help you set up PoinT CLM simulations.

It runs the CLM tools for you to get datasets set up, and copies them to a location you can use them according to the `CLM_USRDATA_NAME` naming convention.

Then it runs `create_newcase` for you and modifies the env settings and namelist appropriately.

PTCLM has a simple ASCII text file for storing basic information for your sites.

We also have complete lists for AmeriFlux and Fluxnet-Canada sites, although we only have the meteorology data for one site.

For other sites you will need to obtain the meteorology data and translate it to a format that the CESM datm model can use.

But, even without meteorology data PTCLM is useful to setup datasets to run with standard `CLM_QIAN` data.

The original authors of PTCLM are: Daniel M. Ricciuto, Dali Wang, Peter E. Thornton, Wilfred M. Post all at Environmental Sciences Division, Oak Ridge National Laboratory (ORNL) and R. Quinn Thomas at Cornell University. It was then modified fairly extensively by Erik Kluzek at NCAR. We want to thank all of these individuals for this contribution to the CESM effort. We also want to thank the folks at University of Michigan Biological Stations (US-UMB) who allowed us to use their Fluxnet station data and import it into our inputdata repository, especially Gil Bohrer the PI on record for this site.

1.7.2 Details of PTCLM

To get help on PTCLM1.110726 use the “–help” option as follows.

```
> cd scripts/ccsm_utils/Tools/lnd/clm/PTCLM
> ./PTCLM.py --help
```

The output to the above command is as follows:

```
Usage: PTCLM.py [options] -d inputdatadir -m machine -s sitename

Python script to create cases to run single point simulations with tower site data.

Options:
  --version           show program's version number and exit
  -h, --help          show this help message and exit

Required Options:
  -d CCSM_INPUT, --csmdata=CCSM_INPUT
                      Location of CCSM input data
  -m MYMACHINE, --machine=MYMACHINE
                      Machine, valid CESM script machine (-m list to list valid
                      machines)
  -s MYSITE, --site=MYSITE
                      Site-code to run, FLUXNET code or CLM1PT name (-s list to list
                      valid names)

Configure and Run Options:
  -c MYCOMPSET, --compset=MYCOMPSET
                      Compset for CCSM simulation (Must be a valid 'I' compset [other
                      than IG compsets], use -c list to list valid compsets)
  --coldstart         Do a coldstart with arbitrary initial conditions
  --caseidprefix=MYCASEID
                      Unique identifier to include as a prefix to the case name
  --cesm_root=BASE_CESM
                      Root CESM directory (top level directory with models and scripts
                      subdirs)
  --debug             Flag to turn on debug mode so won't run, but display what would
                      happen
  --finidat=FINIDAT Name of finidat initial conditions file to start CLM from
  --list              List all valid: sites, compsets, and machines
  --namelist=NAMELIST
                      List of namelist items to add to CLM namelist (example:
                      --namelist="hist_finc1l='TG',hist_ntfrq=-1"
  --QIAN_tower_yrs   Use the QIAN forcing data year that correspond to the tower
                      years
  --rmold             Remove the old case directory before starting
  --run_n=MYRUN_N    Number of time units to run simulation
  --run_units=MYRUN_UNITS
                      Time units to run simulation (steps, days, years, etc.)
  --quiet             Print minimul information on what the script is doing
  --sitegroupname=SITEGROUP
                      Name of the group of sites to search for you selected site in
                      (look for prefix group names in the PTCLM_sitedata directory)
  --stdurbpt          If you want to setup for standard urban namelist settings
  --useQIAN           use QIAN input forcing data instead of tower site meterology
  ↵data
  --verbose           Print out extra information on what the script is doing

Input data generation options:
These are options having to do with generation of input datasets. Note: When
running for supported CLM1PT single-point datasets you can NOT generate new
datasets. For supported CLM1PT single-point datasets, you MUST run with the
following settings: --nopointdata And you must NOT set any of these: --soilgrid
--pftgrid --owritesrf
```

```
--nopointdata          Do NOT make point data (use data already created)
--owritesrf           Overwrite the existing surface datasets if they exist (normally
                      do NOT recreate them)
--pftgrid             Use pft information from global gridded file (rather than site
                      data)
--soilgrid             Use soil information from global gridded file (rather than site
                      data)

Main Script Version Id: $Id: PTCLM.py 47576 2013-05-29 19:11:16Z erik $ Scripts URL:
↪$HeadURL: https://svn-ccsm-models.cgd.ucar.edu/PTCLM/trunk_tags/PTCLM1_130529/PTCLM.
↪py $:
```

Here we give a simple example of using PTCLM1 for a straightforward case of running at the US-UMB Fluxnet site on yellowstone where we already have the meteorology data on the machine. Note, see the Section called [Converting AmeriFlux Data for use by PTCLM](#) for permission information to use this data.

Example 6-1. Example of running PTCLM1 for US-UMB on yellowstone

```
> setenv CSMDATA      $CESMDATAROOT/inputdata
> setenv MYCSMDATA   $HOME/inputdata
> setenv SITE         US-UMB
> setenv MYMACH       yellowstone_intel
> setenv MYCASE        testPTCLM
# First link the standard input files to a location you have write access
> cd scripts
> ./link_dirtree $CSMDATA $MYCSMDATA

# Next build all of the clm tools you will need
> cd ../models/lnd/clm/tools/clm4_5/mksurfdata_map
> gmake
> gmake clean
> cd ../../../../../../../tools/mapping/gen_domain_files/src
> ../../../../../../scripts/ccsm_utils/Machines/configure -mach yellowstone -compiler intel
> gmake
> gmake clean
# next run PTCLM (NOTE -- MAKE SURE python IS IN YOUR PATH)
> cd ../../../../../../scripts/ccsm_utils/Tools/lnd/clm/PTCLM
> ./PTCLM.py -m $MYMACH --case=$MYCASE --site=$SITE --csmdata=$MYCSMDATA --
↪aerdepgrid --ndepgrid
# NOTE: we use --aerdepgrid --ndepgrid so that you use the global
# aerosol and Nitrogen deposition files rather than site-specific ones.
> cd ../../../../../../$MYCASE
# Finally setup, build, and run the case as normal
```

1.7.3 Using PTCLM

There are three types of options to PTCLM1: required, setup/run-time, and dataset generation options. The three required options are the three settings that **MUST** be specified for PTCLM to work at all. The other settings have default values that will default to something useful. The setup/run-time options control how the simulation will be setup and run. The dataset generation options control the generation of datasets needed when PTCLM is run. Most options use a double dash “--” “longname” such as “-list”, but the most common options also have a short-name with a single dash (such as -m instead of -machine).

The required options to PTCLM are: inputdata directory (-d), machine (-m) and site-name (-s). Inputdata directory is the directory where you have the CESM inputdata files, you need to have write access to this directory, so if you are

running on a machine that you do NOT have write access to the standard inputdata location (such as NCAR yellowstone or LBNL hopper) you need to link the standard files to a location you do have control over. We recommend using the scripts/link_dirtree tool to do that. “machine” is the scripts name for the machine/compiler you will be using for your case. And finally site-name is the name of the site that you want to run for. Site-name can either be a valid supported dataset name or a Fluxnet site name from the list of sites you are running on (see the –sitegroupname for more information about the site lists).

After PTCLM is run a case directory where you can then setup, build and run your CESM case as normal. It also creates a README.PTCLM in that directory that documents the commandline options to PTCLM that were used to create it.

After “help” the “list” option is one of the most useful options for getting help on using PTCLM. This option gives you information about some of the other options to PTCLM. To get a list of the machine, sites, and compsets that can be used for PTCLM use the “–list” option as follows.

```
> cd scripts/ccsm_utils/Tools/lnd/clm/PTCLM  
> ./PTCLM.py --list
```

The output to the above command is as follows:

```
/bin/sh: line 1: PTCLM.py: command not found
```

Steps in running PTCLM

1. Setup Inputdata directory with write access (use link_dirtree script)

You need to setup an inputdata directory where you have write access to it. Normally, for NCAR machines the data is on an inputdata where the user does NOT have write access to it. A way that you can get around this is to use the **link_dirtree** script to create softlinks from the normal location to a location you have write access to. So for example on yellowstone:

```
> setenv CSMDATA $CESMDATAROOT/inputdata  
> setenv MYCSMDATA $HOME/inputdata  
> mkdir $MYCSMDATA  
> cd scripts  
> ./link_dirtree $CSMDATA $MYCSMDATA
```

See the Section called [Managing Your Own Data-files in Chapter 3](#) for more information on this.

2. Build the CLM tools Next you need to make sure all the CLM FORTRAN tools are built.

```
> cd models/lnd/clm/tools/clm4_5/mksurfdata_map  
> gmake  
> gmake clean  
> cd ../../../../../../tools/mapping/gen_domain_files/src  
> ../../../../../scripts/ccsm_utils/Machines/configure -mach yellowstone -compiler_=  
intel  
> gmake  
> gmake clean
```

3. Run PTCLM Next you actually run PTCLM1 which does the different things listed below:

(a) PTCLM names your case based on your input

```
[Prefix_]SiteCode_Compset[_QIAN]
```

Where: Prefix is from the caseidprefix option (or blank if not used).

`SiteCode` is the site name you entered with the `-s` option.

`Compset` is the compset name you entered with the `-c` option.

`_QIAN` is part of the name only if the `useQIAN` is used.

For example, the casename for the following will be:

```
> cd scripts
> ./PTCLM.py -m yellowstone_intel -s US-UMB -d $MYCSMDATA -c ICRUCLM45BGC --
→use QIAN "US-UMB_I_2000_CN_QIAN"
```

- (b) PTCLM creates datasets for you It will populate `$MYCSMDATA` with new datasets it creates using the CLM tools.
 - (c) If a transient compset and PTCLM1 finds a `_dynpftdata.txt` file If you are running a transient compset (such as the “`I_1850-2000_CN`” compset) AND you there is a file in the `PTCLM_sitedata` directory under the PTCLM directory called `$SITE_dynaPFTdata.txt` it will use this file for the land-use changes. Otherwise it will leave land-use constant, unless you use the `pftgrid` option so it uses the global dataset for landuse changes. See the Section called Dynamic Land-Use Change Files for use by PTCLM for more information on this. There is a sample transient dataset called `US-Ha1_dynaPFTdata.txt`. Transient compsets, are compsets that create transient land-use change and forcing conditions such as: ‘`I_1850-2000`’, ‘`I_1850-2000_CN`’, ‘`I_RCP8.5_CN`’, ‘`I_RCP6.0_CN`’, ‘`I_RCP4.5_CN`’, or ‘`I_RCP2.6_CN`’.
 - (d) PTCLM creates a pft-physiology for you PTCLM1 will create a local copy of the pft-physiology specific for your site that you could then customize with changes specific for that site.
 - (e) PTCLM creates a `README.PTCLM` for you PTCLM1 will create a simple text file with the command line for it in a file called `README.PTCLM` in the case directory it creates for you.
4. Customize, setup, build and run case as normal You then customize your case as you would normally. See the Chapter 1 chapter for more information on doing this.

PTCLM options

Next we discuss the setup and run-time options, dividing them up into setup, initial condition (IC), and run-time options.

Configure options include:

- `-compset=MYCOMPSET`
- `-caseidprefix=MYCASEID`
- `-cesm_root=BASE_CESM`
- `-namelist=NAMELIST`
- `-rmold`
- `-scratchroot=SCRATCHROOT`
- `-sitegroupname=SITEGROUP`
- `-QIAN_tower_yrs`
- `-useQIAN`

--compset The “`-c`” option is the most commonly used option after the required options, as it specifies the CESM scripts component set to use with PTCLM1. The default compset is the “`ICN`” compset with `CN` on for present day conditions.

--caseidprefix This option gives a prefix to include in the casename when the case is created, in case you want to customize your casenames a bit. By default, casenames are figured out based on the other options. The argument to this option can either be a name to prefix casenames with and/or a pathname to include. Hence, if you want cases to appear in a specific directory you can give the pathname to that directory with this option.

--cesm_root This option is for running PTCLM1 with a different root directory to CESM than the version PTCLM exists in. Normally you do NOT need to use this option.

--namelist This option adds any items given into the CLM user_nl_clm namelist. This allows you to add customizations to the namelist before the clm.buildnml.csh file is created for the case.

--rmold This option will remove an old case directory of the same name if one exists. Otherwise, if an old case directory already exists and you try to run PTCLM it will return with an error.

--scratchroot This option is ONLY valid when using one of the generic machines (the -m option). This passed onto **create_newcase** and gives the location where cases will be built and run.

--sitegroupname In the PTCLM directory there is a subdirectory “PTCLM_sitedata” that contains files with the site, PFT and soil data information for groups of sites. These site groups are all separate ASCII files with the same prefix followed by a “_*data.txt” name. See [the Section called PTCLM Group Site Lists](#) for more information on these files. By default we have provided three different valid group names:

EXAMPLE

AmeriFlux

Fluxnet-Canada

The EXAMPLE is the group used by default and ONLY includes the US-UMB site as that is the only site we have data provided for. The other two site groups include the site information for all of both the AmeriFlux and Fluxnet-Canada sites. You can use the “sitegroupname” option to use one of the other lists, or you can create your own lists using the EXAMPLE file as an example. Your list of sites could be real world locations or could be theoretical “virtual” sites given to exercise CLM on differing biomes for example. Note, see [the Section called Converting AmeriFlux Data for use by PTCLM](#) with permission information to use the US-UMB data.

--useQIAN This option says to use the standard CLM global Qian T62 atmospheric forcing rather than any tower site forcing data available. Otherwise, PTCLM will try to find tower forcing data for the specific site entered.

--QIAN_tower_yrs This option is used with the “useQIAN” option to set the years to cycle over for the Qian data. In this case Qian atmospheric forcing will be used, but the simulation will run over the same years that tower site is available for this site.

IC options include:

- --coldstart
- --finidat=FINIDAT

The coldstart option says to startup with OUT an initial condition file, while the finidat option explicitly gives the initial condition file to use. Obviously, the coldstart and finidat options can NOT be used together.

--coldstart This option ensures that a cold-start will be done with arbitrary initial conditions.

--finidat This option sets the initial condition file to startup the simulation from.

Run-time options include:

- --debug
- --run_n=MYRUN_N
- --run_units=MYRUN_UNITS

- `-stdurbpt`
- `-debug`

This option tells PTCLM to echo what it would do if it were run, but NOT actually run anything. So it will show you the dataset creation commands it would use. It does however, run `create_newcase`, but then it only displays the `xmlchange` commands and changes that it would do. Also note that if you give the “`-rmold`” option it won’t delete the case directory beforehand. Primarily this is intended for debugging the operation of PTCLM.

--run_n This option along with `run_units` is used to set the length for the simulation. “`run_n`” is the number of units to use. The default run length depends on the site, compset, and configuration.

--run_units This option is the units of time to use for the length of the simulation. It is used along with “`run_n`” to set the length of the simulation. The default run length depends on the site, compset, and configuration.

--stdurbpt This option turns on the “`stdurbpt_pd`” use-case for CLM_NML_USE_CASE. This option can NOT be used for compsets that set the use-case to something besides present-day.

The dataset generation options are:

- `-pftgrid`
- `-soilgrid`
- `-nopointdata`
- `-owritesrfaer`

The options that with a “grid” suffix all mean to create datasets using the global gridded information rather than using the site specific point data. By default the site specific point data is used. The “`nopointdata`” and “`owritesrfaer`” options have to do with file creation.

Because supported single-point datasets already have the data created for them, you MUST use the “`nopointdata`” and “`ndegrid`” options when you are using a supported single-point site. You must use “`ndegrid`” even for a compset without CN. You also can NOT use the options: “`soilgrid`”, “`pftgrid`”, “`aerdegrid`”, or “`owritesrfaer`”.

--pftgrid This option says to use the PFT values provided on the global dataset rather than using the specific site based values from the `PTCLM_sitedata/*_pftdata.txt` file when creating the surface dataset. This option must NOT be used when you are using a site that is a supported single point dataset.

--soilgrid This option says to use the soil values provided on the global dataset rather than using the specific site based values from the `PTCLM_sitedata/*_soildata.txt` file when creating the surface dataset. This option must NOT be used when you are using a site that is a supported single point dataset.

--nopointdata This option says to NOT create any input datasets – assume this step has already been done. If datasets weren’t already created, your case will fail when you try to run it. In general the first time you run PTCLM for a new site you want it to generate new datasets, but the next time and future times you want to use this option so that it doesn’t waste a lot of time rebuilding datasets over again.

Note: This option is required when you are using a site that is a supported single point dataset.

--owritesrfaer This option says to overwrite any surface and/or aerosol deposition datasets that were already created. Otherwise, the creation of these files will be skipped if a file is already found (but it WILL create files if they don’t exist). This option must NOT be used when you are using a site that is a supported single point dataset.

1.7.4 Examples of using PTCLM

Now let’s give a few more complex examples using some of the options we have discussed above.

In this first example, we'll demonstrate using a supported single point dataset, which then requires using the “nopointdata”. We'll also demonstrate the compset option, “stdurbpt” and “caseidprefix” options.

Example: Running PTCLM for the Mexicocity supported single point dataset

```
> cd scripts/ccsm_utils/Tools/lnd/clm/PTCLM
> ./PTCLM.py -m yellowstone_intel -s 1x1_mexicocityMEX -d $CSMDATA --nopointdata \
--stdurbpt -c ICRUCLM45 --caseidprefix `pwd`/myPTCLMcases/site
> cd myPTCLMcases/site_1x1_mexicocityMEX_I
> ./cesm_setup
# Now build and run normally
> ./site_1x1_mexicocityMEX_I.build
# Here we show running interactively
> ./site_1x1_mexicocityMEX_I.run
```

Now, let's demonstrate using a different group list, doing a spinup, running with Qian global forcing data, but using tower years to set the years to run over. This uses the options: sitegroupname, useQIAN, and QIANtower_years.

Example: Running PTCLM for a spinup simulation with Qian data for tower years.

```
> cd scripts/ccsm_utils/Tools/lnd/clm/PTCLM
> ./PTCLM.py -m yellowstone_intel -s US-Hal -d $CSMDATA --sitegroupname AmeriFlux --
useQIAN --QIAN_tower_yrs
> cd ../../../../../../US-Hal_ICRUCLM45BGC_QIAN
> ./cesm_setup
# Now build and run normally
> ./US-Hal_ICRUCLM45BGC_QIAN.build
# Here we show running interactively
> ./US-Hal_ICRUCLM45BGC_QIAN.run
...
```

Finally, let's demonstrate using a generic machine (which then requires the scratchroot option), using the global grid for PFT and soil types, and setting the run length to two months.

Example: Running PTCLM on a user-defined machine with global PFT and soil types dataset

```
> cd scripts/ccsm_utils/Tools/lnd/clm/PTCLM
# Note, see the the Section called Converting AmeriFlux Data for use by PTCLM with
# permission information
# to use the US-UMB data.
> ./PTCLM.py -m userdefined_intel -s US-UMB -d $CSMDATA --pftgrid --soilgrid \
--scratchroot $HOME --run_n 2 --run_units nmonths
> cd ../../../../../../US-UMB_ICRUCLM45BGC
# If userdefined is NOT set up for you Uncomment the following and set OS, NTASKS_PER_
# NODE, TMPDIR
# > ./xmlchange OS=$OS,MAX_TASKS_PER_NODE=$NTASKS_PER_NODE,MPILIB=mpi-serial
# > ./xmlchange RUNDIR=$TMPDIR/$USER/\$CASE/run,DIN_LOC_ROOT=$CSMDATA,COMPILER=intel
# > ./xmlchange EXEROOT=$TMPDIR/$USER/\$CASE
> ./cesm_setup
# Now build
> ./US-UMB_ICRUCLM45BGC.userdefined_intel.build
# To get the files from the svn server...
# First list the files from the streams text file
> ../ccsm_utils/Tools/listfilesin_streams \
```

```
-t $HOME/US-UMB_ICRUCLM45BGC/run/clm1PT.1x1pt_US-UMB.stream.txt -l \
> Buildconf/datm.input_data_list
# And now run the script to export data to your machine
> ../ccsm_utils/Tools/check_input_data -export
# Here we show running interactively
> ./US-UMB_ICRUCLM45BGC.userdefined_intel.run
```

1.7.5 Adding PTCLM Site Data

The “sitegroupname” option to PTCLM1.110726 looks for groups of sites in the files in the `PTCLM_sitedata` directory under the `PTCLM` directory. You can add new names available for this option including your own lists of sites, by adding more files in this directory. There are three files for each “sitegroupname”: `$SITEGROUP_sitedata.txt`, `$SITEGROUP_soildata.txt` and `$SITEGROUP_pftdata.txt` (where `$SITEGROUP` is the name that would be entered as “sitegroupname” to `PTCLM`). Each file needs to have the same list of sites, but gives different information: site data, PFT data, and soil data respectively. Although the site codes need to be the same between the three files, the files do NOT have to be in the same order. Each file has a one-line header that lists the contents of each column which are separated by commas. The first column for each of the files is the “site_code” which must be consistent between the three files. The site code can be any unique character string, but in general we use the AmeriFlux site code.

Site data file: “`$SITEGROUP_sitedata.txt`“): The header for this file is:

```
site_code,name,state,lon,lat,elev,startyear,endyear,alignyear
```

The columns: name, state, and elevation are informational only. Name is a longer descriptive name of the site, and state is the state for U.S. sites or country for non U.S. sites. The columns: lon and lat are the longitude and latitude of the location in decimal degrees. The last three columns are the start and ending year for the data and the align year for an 1850 case for the data. The align year is currently unused.

Soil data file: `$SITEGROUP_soildata.txt`): The header for this file is:

```
site_code,soil_depth,n_layers,layer_depth,layer_sand%,layer_clay%
```

The first three fields after “site_code” are currently unused. The only two that are used are the percent sand and clay columns to set the soil texture.

PFT data file: `$SITEGROUP_pftdata.txt``): The header for this file is:

```
site_code,pft_f1,pft_c1,pft_f2,pft_c2,pft_f3,pft_c3,pft_f4,pft_c4,pft_f5,pft_c5
```

This file gives the vegetation coverage for the different vegetation types for the site. The file only supports up to five PFT’s at the same time. The columns with “pft_f” are the fractions for each PFT, and the columns with “pft_c” is the integer index of the given PFT. Look at the pft-physiology file to see what the PFT index for each PFT type is.

Dynamic Land-Use Change Files for use by PTCLM

There is a mechanism for giving site-specific land-use change in `PTCLM`. Adding site specific files to the `PTCLM_sitedata` directory under `PTCLM` allows you to specify the change in vegetation and change in harvesting (for the CN model) for that site. Files are named: `$SITE_dynpftdata.txt`. There is a sample file for the US-Ha1 site called: `US-Ha1_dynpftdata.txt`. The file has a one-line header with the information that the file has, and then one-line for each year with a transition. The header line is as follows:

```
trans_year,pft_f1,pft_c1,pft_f2,pft_c2,pft_f3,pft_c3,pft_f4,pft_c4,pft_f5,pft_c5,har_
↪vh1,har_vh2,har_sh1,har_sh2,har_sh3,graze,hold_harv,hold_graze
```

This file only requires a line for each year where a transition or harvest happens. As in the “pftdata” file above “pft_f” refers to the fraction and “pft_c” refers to the PFT index, and only up to five vegetation types are allowed to co-exist. The last eight columns have to do with harvesting and grazing. The last two columns are whether to hold harvesting and/or grazing constant until the next transition year and will just be either 1 or 0. This file will be converted by the **PTCLM_sitedata/cnrvrt_trnsyrs2_pftdynfile.pl** script in the PTCLM directory to a format that **mksurfdata_map** can read that has an entry for each year for the range of years valid for the compset in question.

Converting AmeriFlux Data for use by PTCLM

AmeriFlux data comes in comma separated format and is available from: <http://public.ornl.gov/ameriflux/dataproducts.shtml>. Before you download the data you need to agree to the usage terms.

Here is a copy of the usage terms from the web-site on June/13/2011.

“The AmeriFlux data provided on this site are freely available and were furnished by individual AmeriFlux scientists who encourage their use. Please kindly inform the appropriate AmeriFlux scientist(s) of how you are using the data and of any publication plans. Please acknowledge the data source as a citation or in the acknowledgments if the data are not yet published. If the AmeriFlux Principal Investigators (PIs) feel that they should be acknowledged or offered participation as authors, they will let you know and we assume that an agreement on such matters will be reached before publishing and/or use of the data for publication. If your work directly competes with the PI’s analysis they may ask that they have the opportunity to submit a manuscript before you submit one that uses unpublished data. In addition, when publishing, please acknowledge the agency that supported the research. Lastly, we kindly request that those publishing papers using AmeriFlux data provide preprints to the PIs providing the data and to the data archive at the Carbon Dioxide Information Analysis Center (CDIAC).”

The above agreement applies to the “US-UMB” dataset imported into our repository as well, and Gil Bohrer is the PI on record for that dataset.

The CESM can NOT handle missing data, so we recommend using the “Level 4” Gap filled datasets. The fields will also need to be renamed. The “WS” column becomes “WIND”, “PREC” becomes “PRECmms”, “RH” stays as “RH”, “TA” becomes “TBOT”, “Rg” becomes “FSDS”, “Rgl” becomes “FLDS”, “PRESS” becomes “PSRF”. “ZBOT” can just be set to the constant of “30” (m). The units of Temperature need to be converted from “Celsius” to “Kelvin” (use the value in SHR_CONST_TKFRZ in the file models/csm_share/shr/shr_const.F90 of 273.15. The units of Pressure also need to be converted from “kPa” to “Pa”. LATIXY, and LONGXY should also be set to the latitude and longitude of the site.

Example: PTCLM transient example over a shorter time period

This is an example of using PTCLM for Harvard Forest (AmeriFlux site code US-Ha1) for transient land use 1991-2006. In order to do this we would’ve needed to have converted the AmeriFlux data into NetCDF format as show in the the Section called [Converting AmeriFlux Data for use by PTCLM](#) section above. Also note that this site has a site-specific dynamic land-use change file for it PTCLM_sitedata/US-Ha1_dynpftdata.txt in the PTCLM directory and this file will be used for land-use change and harvesting rather than the global dataset.

```
> cd scripts/ccsm_utils/Tools/lnd/clm/PTCLM
# We are going to use forcing data over 1991 to 2006, but we need to start with
# a transient compset to do so, so we use the 20th Century transient: 1850-2000
# Note: When creating the fpftdyn dataset for this site it will use the
#       PTCLM_sitedata/US-Ha1_dynpftdata.txt
# file for land-use change and harvesting
> ./PTCLM.py -m yellowstone_intel -s US-Ha1 -d $MYCSMDATA --sitegroupname AmeriFlux -
  ↵c I20TRCRUCLM45BGC
> mkdir $MYCSMDATA/atm/datm7/CLM1PT_data/1x1pt_US-Ha1
> cd $MYCSMDATA/atm/datm7/CLM1PT_data/1x1pt_US-Ha1
# Copy data in NetCDF format to this directory, filenames should be YYYY-MM.nc
```

```

# The fieldnames on the file should be:
#   FLDS,FSDS,LATIXY,    LONGXY,    PRECTmms,PSRF,RH,TBOT,WIND,ZBOT
# With units
#   W/m2,W/m2,degrees_N,degrees_E,mm/s,    Pa, %, K, m/s, m
# The time coordinate units should be: days since YYYY-MM-DD 00:00:00
> cd ../../../../../US-Ha1_I20TRCRUCLM45BGC
# Now we need to set the start date to 1991, and make sure the align year is for 1991
> ./xmlchange RUN_STARTDATE=1991-01-01,DATM_CLMNCEP_YR_ALIGN=1991
# Similarly for Nitrogen deposition data we cycle over: 1991 to 2006
> cat << EOF >> user_nl_clm
model_year_align_ndep=1991,stream_year_first_ndep=1991,stream_year_last_ndep=2006
EOF
# Now setup the case, and we'll edit the datm namelist for prescribed aerosols
> ./cesm_setup

```

1.8 Troubleshooting

1.8.1 Trouble Shooting

In this chapter we give some guidance on what to do when you encounter some of the most common problems. We can't cover all the problems that a user could potentially have, but we will try to help you recognize some of the most common situations. And we'll give you some suggestions on how to approach the problem to come up with a solution.

In general you will run into one of three type of problems:

1. *setup-time*
2. *build-time*
3. *run-time*

Setup Problems

The first type of problem happens when you invoke the **case.setup** command. This indicates there is something wrong with your input datasets, or the details of what you are trying to setup the model to do. There's also a trouble-shooting chapter in the [CESM1.2.0 Scripts User's Guide](#). Many of the problems with configuration can be resolved with the guidelines given there. Here we will restrict ourselves to problems from the input files.

Example: Missing datasets

```

> ./create_newcase -case ne60rcp6 -res ne60_g16 -compset IRCP60CN \
-mach yellowstone_intel
> ./case.setup

```

The following is what is displayed to the screen.

```

.
.
.
Running preview_namelist script
CLM configure done.
CLM adding use_case 1850-2100_rcp6_transient defaults for var clm_demand with val_
→fpftdyn

```

```
CLM adding use_case 1850-2100_rcp6_transient defaults for var clm_start_type with val_
↪startup
CLM adding use_case 1850-2100_rcp6_transient defaults for var model_year_align_ndep_
↪with val 1850
CLM adding use_case 1850-2100_rcp6_transient defaults for var rcp with val 6
CLM adding use_case 1850-2100_rcp6_transient defaults for var sim_year with val 1850
CLM adding use_case 1850-2100_rcp6_transient defaults for var sim_year_range with val_
↪1850-2100
CLM adding use_case 1850-2100_rcp6_transient defaults for var stream_year_first_ndep_
↪with val 1850
CLM adding use_case 1850-2100_rcp6_transient defaults for var stream_year_last_ndep_
↪with val 2100
CLM adding use_case 1850-2100_rcp6_transient defaults for var use_case_desc with val_
↪Simulate transient land-use, aerosol and Nitrogen deposition changes
with historical data from 1850 to 2005 and then with the RCP6 scenario from AIM

build-namelist - No default value found for fpftdyn.
    Are defaults provided for this resolution and land mask?
ERROR: clm.buildnml.csh failed
ERROR: /Users/erik/clm_cesm1_1_1_rel/scripts/ne60rcp6/preview_namelists failed: 25344
```

The important thing to note here is the line:

```
ERROR: clm.buildnml.csh failed
```

which tells us that the problem is in the land **clm.buildnml.csh**. It may also indicate problems in one of the other buildnml.csh files (atm, cesm, cpl, glc, ice, or ocn), in which case you should consult the appropriate model user's guide.

In the example, the error is that the CLM XML database does NOT have a finidat for the given resolution, rcp scenario and ocean mask. That means you will need to create the file and then supply the file into your case. See [Chapter 2](#) for more information on creating files, and see [Chapter 3](#) for more information on adding files to the XML database. Alternatively, you can provide the file to your case by creating a user namelist as shown in [the Section called User Namelist in Chapter 1](#).

Note: The two most common problems from your **clm.buildnml.csh** will be errors from the CLM **configure** or **build-namelist**. For more information on these scripts see: [the Section called More information on the CLM configure script in Chapter 1](#) and [the section on CLM_BLDNML_OPTS](#).

Build problems

The following is an example of running the build for a case and having it fail in the land model build. As you can see it lists which model component is being built and the build log for that component.

```
CCSM BUILDEXE SCRIPT STARTING
- Build Libraries: mct pio csm_share
Sat Jun 19 21:21:19 MDT 2010 /ptmp/erik/test_build/mct/mct.bldlog.100619-212107
Sat Jun 19 21:22:18 MDT 2010 /ptmp/erik/test_build/pio/pio.bldlog.100619-212107
Sat Jun 19 21:23:18 MDT 2010
/ptmp/erik/test_build/csm_share/csm_share.bldlog.100619-212107
Sat Jun 19 21:24:00 MDT 2010 /ptmp/erik/test_build/run/cpl.bldlog.100619-212107
Sat Jun 19 21:24:00 MDT 2010 /ptmp/erik/test_build/run/atm.bldlog.100619-212107
Sat Jun 19 21:24:06 MDT 2010 /ptmp/erik/test_build/run/lnd.bldlog.100619-212107
```

```
ERROR: clm.buildexe.csh failed, see /ptmp/erik/test_build/run/lnd.bldlog.100619-
↪212107
ERROR: cat /ptmp/erik/test_build/run/lnd.bldlog.100619-212107
```

You can then examine the build log that failed and see what went wrong. Most compilers will give the full filepath and line number for the file that failed to compile.

Run Time Problems

Tracking down problems while the model is running is much more difficult to do than setup or build problems. In this section we will give some suggestions on how to find run time problems. Below we show the log file results of a job that aborted while running.

```
CCSM PRESTAGE SCRIPT HAS FINISHED SUCCESSFULLY
Sun Jun 20 18:24:06 MDT 2010 -- CSM EXECUTION BEGINS HERE
Sun Jun 20 18:24:35 MDT 2010 -- CSM EXECUTION HAS FINISHED
Model did not complete - see /ptmp/erik/test_run/run/cpl.log.100620-182358
```

In the next section we will talk about using the different log files to track down problems, and find out where the problem is coming from. In the section after that we give some general advice on debugging problems and some suggestions on ideas that may be helpful to track the problem down. Some of the examples below are from the `models/lnd/clm/doc/KnownBugs` file.

Tracking Problems by Querying Log Files

The first thing to do when tracking down problems is to query the different log files to see if you can discover where the problem occurs, and any error messages about it. It's important to figure out if the problem comes in at initialization or in the run phase of the model, and in which model component the problem happens. There are different log files for the different major components, and they all end with the date and time in YYMMDD-HHMMSS format (2-digit: year, month, day, hour minute and second). When the model runs to completion the log files will be copied to the logs directory in the script directory, but when the model fails they will remain in the run directory. Here's an example list of log files from an "I" case where the model dies in the land model initialization. For "I" cases the sea-ice and ocean components are just stubs and don't create log files (and unless running with the active land-ice model "glc" log files won't be created either).

```
atm.log.100620-182358
cesm.log.100620-182358
cpl.log.100620-182358
lnd.log.100620-182358
```

The coupler log file

The first log file to check is the coupler log file so that you can see where the model dies and which model component it fails in. When the model dies at initialization the last model component listed is the component that failed.

Example of a case that fails in the CLM land model initialization.

```
(seq_timemgr_clockPrint)      Prev Time    = 00001201  00000
(seq_timemgr_clockPrint)      Next Time    = 99991201  00000
(seq_timemgr_clockPrint)      Intervl yms =      9999          0
                               0
(seq_mct_drv) : Initialize each component: atm, lnd, ocn, and ice
```

```
(seq_mct_drv) : Initialize atm component
(seq_mct_drv) : Initialize lnd component
```

The cesm log file

The cesm log files are to some extent the “garbage collection” of log output. The CLM sends it’s output from it’s master processor, but sends other output and possibly errors to the cesm log file. Because, of this, often error messages are somewhere in the cesm log file. However, since there is so much other output it may be difficult to find. For example, here is some output from an older version of CESM (CESM1.0.2) where the RTM river routing file (before it was converted to NetCDF) was not provided and the error on the open statement for the file was embedded near the end of the cesm log file.

```
NODE# NAME
( 0) be1105en.ucar.edu
"/gpfs/proj2/fis/cgd/home/erik/clm_trunk/models/lnd/clm/src/riverroute/RtmMod.F90",_
↪line
239: 1525-155 The file name provided in the OPEN statement for unit 1 has zero length_
↪or
contains all blanks. The program will recover by ignoring the OPEN statement.
"/gpfs/proj2/fis/cgd/home/erik/clm_trunk/models/lnd/clm/src/riverroute/RtmMod.F90",_
↪line
241: 1525-001 The READ statement on the file fort.1 cannot be completed because the_
↪end
of the file was reached. The program will stop.

Running: ./cesm.exe
Please wait...

Memory usage for ./cesm.exe (task # 0) is:      51696 KB. Exit status: 1. Signal:_
↪0
```

Although the example is from an earlier version of the model it still serves to illustrate finding problems from the cesm log file.

When working with the cesm log file, for a run-time problem, you will need to be able to separate it’s output into three categories: pre-crash, crash, and post-crash. The pre-crash section is everything that is normal output for good operation of the model. The crash section is the section where the model dies and reports on the actual problem. the post-crash section is the cleanup and finalization after the model dies. The most important part of this of course is the crash section. The tricky part is distinguishing it from the other sections. Also because the cesm log file most likely has duplicated output from multiple processors it is even more difficult to distinguish the different sections and to some extent the sections may be intertwined, as different processors reach the different sections at different times. Because, of this reducing the number of processors for your simulation may help you sort out the output in the file (see the [Section called Run with a smaller set of processors](#)). Also much of the output from the cesm log file are system level information having to do with MPI multiprocessing. Usually you can ignore this information, but it makes it more difficult to trudge through.

Sometimes the cesm log file is the ONLY file available, because the model terminates early in initialization. In this case understanding the output in the cesm log file becomes even more important. This also indicates the model did NOT advance far enough to reach the initialization of the individual model components. This may mean that the initialization of the multiprocessing for MPI and/or OpenMP failed, or that the reading of the driver namelist file “drv_in” failed.

Here we show those three sections for a cesm log file where a two task job failed on reading the namelist file. For a typical job with many tasks similar sections of this will be repeated not just twice but for each task and hence make it harder to read.

Pre-crash section of the cesm log file

```

ATTENTION: 0031-386 MP_INSTANCES setting ignored when LoadLeveler is not being used.

ATTENTION: 0031-386 MP_INSTANCES setting ignored when LoadLeveler is not being used.
ATTENTION: 0031-378 MP_EUIDEVICE setting ignored when LoadLeveler is not being used.
ATTENTION: 0031-386 MP_INSTANCES setting ignored when LoadLeveler is not being used.
  0:INFO: 0031-724 Executing program: </usr/local/lsf/7.0/aix5-64/bin/lsnrt_run>
  1:INFO: 0031-724 Executing program: </usr/local/lsf/7.0/aix5-64/bin/lsnrt_run>
  0:/contrib/bin/cesm_launch: process 401894 bound to logical CPU 0 on host be0310en.
→ucar.edu ...
  1:/contrib/bin/cesm_launch: process 439264 bound to logical CPU 1 on host be0310en.
→ucar.edu ...
  0:INFO: 0031-619 64bit(us, Packet striping on) ppe_rmas MPC1_MSG: MPI/MPC1_
→library was compiled on Wed Aug  5 13:36:06 2009
  0:
  1:LAPI version #14.26 2008/11/23 11:02:30 1.296 src/rsct/lapi/lapi.c, lapi, rsct_
→rpt53, rpt53s004a 09/04/29 64bit(us) library compiled on Wed Apr 29 15:30:42 2009
  1:.
  1:LAPI is using lightweight lock.
  0:LAPI version #14.26 2008/11/23 11:02:30 1.296 src/rsct/lapi/lapi.c, lapi, rsct_
→rpt53, rpt53s004a 09/04/29 64bit(us) library compiled on Wed Apr 29 15:30:42 2009
  0:.
  0:LAPI is using lightweight lock.
  0:Use health ping for failover/recovery
  1:Use health ping for failover/recovery
  0:Initial communication over instance 2.
  1:Initial communication over instance 0.
  1:IB RDMA initialization completed successfully
  1:The MPI shared memory protocol is used for the job
  0:IB RDMA initialization completed successfully
  0:LAPI job ID for this job is: 1684890719
  0:The MPI shared memory protocol is used for the job
  0:(seq_comm_setcomm) initialize ID ( 7 GLOBAL ) pelist = 0 1 1 (_
→npes = 2) ( nthreads = 1)
  0:(seq_comm_setcomm) initialize ID ( 2 ATM ) pelist = 0 1 1 (_
→npes = 2) ( nthreads = 1)
  0:(seq_comm_setcomm) initialize ID ( 1 LND ) pelist = 0 1 1 (_
→npes = 2) ( nthreads = 1)
  0:(seq_comm_setcomm) initialize ID ( 4 ICE ) pelist = 0 1 1 (_
→npes = 2) ( nthreads = 1)
  0:(seq_comm_setcomm) initialize ID ( 5 GLC ) pelist = 0 1 1 (_
→npes = 2) ( nthreads = 1)
  0:(seq_comm_setcomm) initialize ID ( 3 OCN ) pelist = 0 1 1 (_
→npes = 2) ( nthreads = 1)
  0:(seq_comm_setcomm) initialize ID ( 6 CPL ) pelist = 0 1 1 (_
→npes = 2) ( nthreads = 1)
  0:(seq_comm_joincomm) initialize ID ( 8 CPLATM ) join IDs = 6 2 (_
→npes = 2) ( nthreads = 1)
  0:(seq_comm_joincomm) initialize ID ( 9 CPLLND ) join IDs = 6 1 (_
→npes = 2) ( nthreads = 1)
  0:(seq_comm_joincomm) initialize ID ( 10 CPLICE ) join IDs = 6 4 (_
→npes = 2) ( nthreads = 1)
  0:(seq_comm_joincomm) initialize ID ( 11 CPLOCN ) join IDs = 6 3 (_
→npes = 2) ( nthreads = 1)
  0:(seq_comm_joincomm) initialize ID ( 12 CPLGLC ) join IDs = 6 5 (_
→npes = 2) ( nthreads = 1)
  0:
  0: (seq_comm_printcomms) ID layout : global pes vs local pe for each ID

```

```

0:      gpe          LND          ATM          OCN          ICE          GLC          CPL          GLOBAL
→CPLATM    CPLLND    CPLICE    CPLOCN    CPLGLC    nthrds
0:      ---          ----          ----          ----          ----          ----          ----          ----          ----
→--      ----          ----          ----          ----          ----          ----          ----          ----          ----
0:      0   :     0   0   0   0   0   0   0   0   0
→ 0   0   0   0   0   0   1   1   1   1
1:      1   :     1   1   1   1   1   1   1   1   1
→ 1   1   1   1   1   1   1   1   1   1
1:
0: (t_initf) Read in prof_inparm namelist from: drv_in
1: (seq_io_init) cpl_io_stride, iotasks or root out of bounds - resetting to
→defaults 4 0 1
0: piolib_mod.f90 1353 1 2 1 2
1: piolib_mod.f90 1353 1 2 1 2
0: pio_support::pio_die:: myrank= 0 : ERROR: piolib_mod.f90: 1354 : not enough
→procs for the stride
1: pio_support::pio_die:: myrank= 1 : ERROR: piolib_mod.f90: 1354 : not enough
→procs for the stride

```

Crash section of the cesm log file

```

0:
0: Traceback:
1:
1: Traceback:
0:   Offset 0x00000c4c in procedure __pio_support_NMOP_piodie, near line 88 in file
→pio_support.F90.in
1:   Offset 0x00000c4c in procedure __pio_support_NMOP_piodie, near line 88 in file
→pio_support.F90.in
0:   Offset 0x00000fd0 in procedure __piolib_mod_NMOP_init, near line 1354 in file
→piolib_mod.F90
1:   Offset 0x00000fd0 in procedure __piolib_mod_NMOP_init, near line 1354 in file
→piolib_mod.F90
1:   Offset 0x00000398 in procedure __seq_io_mod_NMOP_seq_io_init, near line 247 in
→file /gpfs/proj2/fis/cgd/home/erik/clm_trunk/models/drv/shr/seq_io_mod.F90
0:   Offset 0x00000398 in procedure __seq_io_mod_NMOP_seq_io_init, near line 247 in
→file /gpfs/proj2/fis/cgd/home/erik/clm_trunk/models/drv/shr/seq_io_mod.F90
0:   Offset 0x0001aa88 in procedure ccsm_driver, near line 465 in file /gpfs/proj2/
→fis/cgd/home/erik/clm_trunk/models/drv/driver/ccsm_driver.F90
0:   --- End of call chain ---
1:   Offset 0x0001aa88 in procedure ccsm_driver, near line 465 in file /gpfs/proj2/
→fis/cgd/home/erik/clm_trunk/models/drv/driver/ccsm_driver.F90
1:   --- End of call chain ---

```

Post-crash section of the cesm log file

```

1:Communication statistics of task 1 is associated with task key: 1684890719_1
0:Communication statistics of task 0 is associated with task key: 1684890719_0
0:
0:Running: ./cesm.exe
0:Please wait...
0:
0:Memory usage for ./cesm.exe (task # 0) is: 198892 KB. Exit status: 134.
→Signal: 0
1:
1:Running: ./cesm.exe
1:Please wait...
1:

```

```

1:Memory usage for ./cesm.exe (task # 0) is: 198572 KB. Exit status: 134.
→Signal: 0
INFO: 0031-656 I/O file STDOUT closed by task 0
INFO: 0031-656 I/O file STDERR closed by task 0
ERROR: 0031-250 task 0: IOT/Abort trap
INFO: 0031-656 I/O file STDOUT closed by task 1
INFO: 0031-656 I/O file STDERR closed by task 1
ERROR: 0031-250 task 1: IOT/Abort trap
INFO: 0031-639 Exit status from pm_respond = 0
ATTENTION: 0031-386 MP_INSTANCES setting ignored when LoadLeveler is not being used.
Job /usr/local/lsf/7.0/aix5-64/bin/poejob /contrib/bin/ccsm_launch /contrib/bin/job_
→memusage.exe ./cesm.exe

TID HOST_NAME COMMAND_LINE STATUS TERMINATION_TIME
===== ===== ===== ===== =====
00000 be0310en /contrib/bin/ccs Exit (134) 08/31/2010 12:32:57
00001 be0310en /contrib/bin/ccs Exit (134) 08/31/2010 12:32:57

```

The CLM log file

Of course when you are working with and making changes to CLM, most of your focus will be on the CLM log file and the errors it shows. As already pointed out if you don't see errors in the lnd.log.* file you should look in the cesm.log.* to see if any errors showed up there.

Here's an example of the lnd.log.* file when running PTS_MODE with initial conditions (this is bug 1025 in the models/lnd/clm/doc/KnownLimitations file).

```

Successfully initialized variables for accumulation

reading restart file I2000CN_f09_g16_c100503.clm2.r.0001-01-01-00000.nc
Reading restart dataset
ERROR - setlatlon.F:Cant get variable dim for lat or lsmlat
ENDRUN: called without a message string

```

The DATM log file

When working with "I cases" the second most common problems after CLM problems are problems with the data atmosphere model. So examining the atm.log.* is important.

Here's an example of a problem that occurs when the wrong prescribed aerosol file is given to a pt1_pt1 simulation.

```

(datm_comp_init) atm mode = CLMNCEP
(shr_strdata_init) calling shr_dmodel_mapSet for fill
(shr_strdata_init) calling shr_dmodel_mapSet for remap
('shr_map_getWts') ERROR: yd outside bounds 19.5000000000000000000000
(shr_sys_abort) ERROR: ('shr_map_getWts') ERROR yd outside 90 degree bounds
(shr_sys_abort) WARNING: calling shr_mpi_abort() and stopping

```

The batch log files

The names of the batch log files will depend on the batch system of the machine that is being used. They will normally be in the script directory. Usually, they don't contain important information, but they are a last resort place to look

for error messages. On the NCAR system “yellowstone” the batch files are called with names that start with the batch submission script and then either “stderr.o” or “stdout.o”, with the job number at the end.

General Advice on Debugging Run time Problems

Here are some suggestions on how to track down a problem while running. In general if the problem still occurs for a simpler case, it will be easier to track down.

1. *Run in DEBUG mode*
2. *Run with a smaller set of processors*
3. *Run in serial mode with a single processor*
4. *Run at a lower resolution*
5. *Run a simpler case*
6. *Run with a debugger*

Run in DEBUG mode

The first thing to try is to run in DEBUG mode so that float point trapping will be triggered as well as array bounds checking and other things the compiler can turn on to help you find problems. To do this edit the `env_build.xml` file and set DEBUG to TRUE as follows:

```
> ./xmlchange DEBUG=TRUE
```

Run with a smaller set of processors

Another way to simplify the system is to run with a smaller set of processors. You will need to clean the setup and edit the `-env_mach_pes.xml`. For example, to run with four processors:

```
> ./case.setup -clean
> ./xmlchange NTASKS_ATM=4,NTASKS_LND=4,NTASKS_ICE=4,NTASKS_OCN=4,NTASKS_CPL=4,NTASKS_
  ↵GLC=4
> ./case.setup
```

Another recommended simplification is to run without threading, so set the NTHRDS for each component to “1” if it isn’t already. Sometimes, multiprocessing problems require a certain number of processors before they occur so you may not be able to debug the problem without enough processors. But, it’s always good to reduce it to as low a number as possible to make it simpler. For threading problems you may have to have threading enabled to find the problem, but you can run with 1, 2, or 3 threads to see what happens.

Run in serial mode with a single processor

Simplifying to one processor removes all multi-processing problems and makes the case as simple as possible. If you can enable `MPILIB=mpi-serial` you will also be able to run interactively rather than having to submit to a job queue, which sometimes makes it easier to run and debug. If you can use `MPILIB=mpi-serial` you can also use threading, but still run interactively in order to use more processors to make it faster if needed.

```

> ./case.setup -clean
# Set tasks and threads for each component to 1
# You could also set threads to something > 1 for speed, but still
# run interactively if threading isn't an issue.

> ./xmlchange NTASKS_ATM=1,NTHRDS_ATM=1,NTASKS_LND=1,NTHRDS_LND=1,NTASKS_ICE=1,NTHRDS_
  ↪ICE=1
> ./xmlchange NTASKS_OCN=1,NTHRDS_OCN=1,NTASKS_CPL=1,NTHRDS_CPL=1,NTASKS_GLC=1,NTHRDS_
  ↪GLC=1
# set MPILIB to mpi-serial so that you can run interactively
> ./xmlchange MPILIB=mpi-serial
> ./case.setup
# Then build your case
# And finally run, by running the *.run script interactively

```

Run at a lower resolution

If you can create a new case running at a lower resolution and replicate the problem it may be easier to solve. This of course requires creating a whole new case, and trying out different lower resolutions.

Run a simpler case

Along the same lines, you might try running a simpler case, trying another compset with a simpler setup and see if you can replicate the problem and then debug from that simpler case. Again, of course you will need to create new cases to do this.

Run with a debugger

Another suggestion is to run the model with a debugger such as: **dbx**, **gdb**, or **totalview**. Often to run with a debugger you will need to reduce the number of processors as outlined above. Some debuggers such as **dbx** will only work with one processor, while more advanced debuggers such as **totalview** can work with both MPI tasks and OMP threads. Even simple debuggers though can be used to query core files, to see where the code was at when it died (for example using the **where** in **dbx** for a core file can be very helpful). For help in running with a debugger you will need to contact your system administrators for the machine you are running on.

CHAPTER 2

CLM TECHNICAL NOTE

July 2017

Technical Description of version 5.0 of the Community Land Model (CLM)

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2.1 Introduction

The purpose of this document is to fully describe the biogeophysical and biogeochemical parameterizations and numerical implementation of version 5.0 of the Community Land Model (CLM5.0). Scientific justification and evaluation of these parameterizations can be found in the referenced scientific papers (`rst_References`). This document and the CLM5.0 User's Guide together provide the user with the scientific description and operating instructions for CLM.

2.1.1 Model History

Inception of CLM

The early development of the Community Land Model can be described as the merging of a community-developed land model focusing on biogeophysics and a concurrent effort at NCAR to expand the NCAR Land Surface Model (NCAR LSM, [Bonan 1996](#)) to include the carbon cycle, vegetation dynamics, and river routing. The concept of a community-developed land component of the Community Climate System Model (CCSM) was initially proposed at the CCSM Land Model Working Group (LMWG) meeting in February 1996. Initial software specifications and development focused on evaluating the best features of three existing land models: the NCAR LSM ([Bonan 1996, 1998](#)) used in the Community Climate Model (CCM3) and the initial version of CCSM; the Institute of Atmospheric Physics, Chinese Academy of Sciences land model (IAP94) ([Dai and Zeng 1997](#)); and the Biosphere-Atmosphere Transfer Scheme (BATS) ([Dickinson et al. 1993](#)) used with CCM2. A scientific steering committee was formed to review the initial specifications of the design provided by Robert Dickinson, Gordon Bonan, Xubin Zeng, and Yongjiu Dai and to facilitate further development. Steering committee members were selected so as to provide guidance and expertise in disciplines not generally well-represented in land surface models (e.g., carbon cycling, ecological modeling, hydrology, and river routing) and included scientists from NCAR, the university community, and government laboratories (R. Dickinson, G. Bonan, X. Zeng, Paul Dirmeyer, Jay Famiglietti, Jon Foley, and Paul Houser).

The specifications for the new model, designated the Common Land Model, were discussed and agreed upon at the June 1998 CCSM Workshop LMWG meeting. An initial code was developed by Y. Dai and was examined in March 1999 by Mike Bosilovich, P. Dirmeyer, and P. Houser. At this point an extensive period of code testing was initiated. Keith Oleson, Y. Dai, Adam Schlosser, and P. Houser presented preliminary results of offline 1-dimensional testing at the June 1999 CCSM Workshop LMWG meeting. Results from more extensive offline testing at plot, catchment, and large scale (up to global) were presented by Y. Dai, A. Schlosser, K. Oleson, M. Bosilovich, Zong-Liang Yang, Ian Baker, P. Houser, and P. Dirmeyer at the LMWG meeting hosted by COLA (Center for Ocean-Land-Atmosphere Studies) in November 1999. Field data used for validation included sites adopted by the Project for Intercomparison of Land-surface Parameterization Schemes ([Henderson-Sellers et al. 1993](#)) (Cabauw, Valdai, Red-Arkansas river basin) and others [FIFE ([Sellers et al. 1988](#)), BOREAS ([Sellers et al. 1995](#)), HAPEX-MOBILHY ([André et al. 1986](#)), ABRACOS ([Gash et al. 1996](#)), Sonoran Desert ([Unland et al. 1996](#)), GSWP ([Dirmeyer et al. 1999](#))]. Y. Dai also presented results from a preliminary coupling of the Common Land Model to CCM3, indicating that the land model could be successfully coupled to a climate model.

Results of coupled simulations using CCM3 and the Common Land Model were presented by X. Zeng at the June 2000 CCSM Workshop LMWG meeting. Comparisons with the NCAR LSM and observations indicated major improvements to the seasonality of runoff, substantial reduction of a summer cold bias, and snow depth. Some deficiencies related to runoff and albedo were noted, however, that were subsequently addressed. Z.-L. Yang and I. Baker demonstrated improvements in the simulation of snow and soil temperatures. Sam Levis reported on efforts to incorporate a river routing model to deliver runoff to the ocean model in CCSM. Soon after the workshop, the code was delivered to

NCAR for implementation into the CCSM framework. Documentation for the Common Land Model is provided by [Dai et al. \(2001\)](#) while the coupling with CCM3 is described in [Zeng et al. \(2002\)](#). The model was introduced to the modeling community in [Dai et al. \(2003\)](#).

CLM2

Concurrent with the development of the Common Land Model, the NCAR LSM was undergoing further development at NCAR in the areas of carbon cycling, vegetation dynamics, and river routing. The preservation of these advancements necessitated several modifications to the Common Land Model. The biome-type land cover classification scheme was replaced with a plant functional type (PFT) representation with the specification of PFTs and leaf area index from satellite data ([Oleson and Bonan 2000](#); [Bonan et al. 2002a, b](#)). This also required modifications to parameterizations for vegetation albedo and vertical burying of vegetation by snow. Changes were made to canopy scaling, leaf physiology, and soil water limitations on photosynthesis to resolve deficiencies indicated by the coupling to a dynamic vegetation model. Vertical heterogeneity in soil texture was implemented to improve coupling with a dust emission model. A river routing model was incorporated to improve the fresh water balance over oceans. Numerous modest changes were made to the parameterizations to conform to the strict energy and water balance requirements of CCSM. Further substantial software development was also required to meet coding standards. The resulting model was adopted in May 2002 as the Community Land Model (CLM2) for use with the Community Atmosphere Model (CAM2, the successor to CCM3) and version 2 of the Community Climate System Model (CCSM2).

K. Oleson reported on initial results from a coupling of CCM3 with CLM2 at the June 2001 CCSM Workshop LMWG meeting. Generally, the CLM2 preserved most of the improvements seen in the Common Land Model, particularly with respect to surface air temperature, runoff, and snow. These simulations are documented in [Bonan et al. \(2002a\)](#). Further small improvements to the biogeophysical parameterizations, ongoing software development, and extensive analysis and validation within CAM2 and CCSM2 culminated in the release of CLM2 to the community in May 2002.

Following this release, Peter Thornton implemented changes to the model structure required to represent carbon and nitrogen cycling in the model. This involved changing data structures from a single vector of spatially independent sub-grid patches to one that recognizes three hierarchical scales within a model grid cell: land unit, snow/soil column, and PFT. Furthermore, as an option, the model can be configured so that PFTs can share a single soil column and thus “compete” for water. This version of the model (CLM2.1) was released to the community in February 2003. CLM2.1, without the compete option turned on, produced only round off level changes when compared to CLM2.

CLM3

CLM3 implemented further software improvements related to performance and model output, a re-writing of the code to support vector-based computational platforms, and improvements in biogeophysical parameterizations to correct deficiencies in the coupled model climate. Of these parameterization improvements, two were shown to have a noticeable impact on simulated climate. A variable aerodynamic resistance for heat/moisture transfer from ground to canopy air that depends on canopy density was implemented. This reduced unrealistically high surface temperatures in semi-arid regions. The second improvement added stability corrections to the diagnostic 2-m air temperature calculation which reduced biases in this temperature. Competition between PFTs for water, in which PFTs share a single soil column, is the default mode of operation in this model version. CLM3 was released to the community in June 2004. [Dickinson et al. \(2006\)](#) describe the climate statistics of CLM3 when coupled to CCSM3.0. [Hack et al. \(2006\)](#) provide an analysis of selected features of the land hydrological cycle. [Lawrence et al. \(2007\)](#) examine the impact of changes in CLM3 hydrological parameterizations on partitioning of evapotranspiration (ET) and its effect on the timescales of ET response to precipitation events, interseasonal soil moisture storage, soil moisture memory, and land-atmosphere coupling. [Qian et al. \(2006\)](#) evaluate CLM3’s performance in simulating soil moisture content, runoff, and river discharge when forced by observed precipitation, temperature and other atmospheric data.

CLM3.5

Although the simulation of land surface climate by CLM3 was in many ways adequate, most of the unsatisfactory aspects of the simulated climate noted by the above studies could be traced directly to deficiencies in simulation of the hydrological cycle. In 2004, a project was initiated to improve the hydrology in CLM3 as part of the development of CLM version 3.5. A selected set of promising approaches to alleviating the hydrologic biases in CLM3 were tested and implemented. These included new surface datasets based on Moderate Resolution Imaging Spectroradiometer (MODIS) products, new parameterizations for canopy integration, canopy interception, frozen soil, soil water availability, and soil evaporation, a TOPMODEL-based model for surface and subsurface runoff, a groundwater model for determining water table depth, and the introduction of a factor to simulate nitrogen limitation on plant productivity. [Oleson et al. \(2008a\)](#) show that CLM3.5 exhibits significant improvements over CLM3 in its partitioning of global ET which result in wetter soils, less plant water stress, increased transpiration and photosynthesis, and an improved annual cycle of total water storage. Phase and amplitude of the runoff annual cycle is generally improved. Dramatic improvements in vegetation biogeography result when CLM3.5 is coupled to a dynamic global vegetation model. [Stöckli et al. \(2008\)](#) examine the performance of CLM3.5 at local scales by making use of a network of long-term ground-based ecosystem observations [FLUXNET ([Balocchi et al. 2001](#))]. Data from 15 FLUXNET sites were used to demonstrate significantly improved soil hydrology and energy partitioning in CLM3.5. CLM3.5 was released to the community in May, 2007.

CLM4

The motivation for the next version of the model, CLM4, was to incorporate several recent scientific advances in the understanding and representation of land surface processes, expand model capabilities, and improve surface and atmospheric forcing datasets ([Lawrence et al. 2011](#)). Included in the first category are more sophisticated representations of soil hydrology and snow processes. In particular, new treatments of soil column-groundwater interactions, soil evaporation, aerodynamic parameters for sparse/dense canopies, vertical burial of vegetation by snow, snow cover fraction and aging, black carbon and dust deposition, and vertical distribution of solar energy for snow were implemented. Major new capabilities in the model include a representation of the carbon-nitrogen cycle (CLM4CN, see next paragraph for additional information), the ability to model land cover change in a transient mode, inclusion of organic soil and deep soil into the existing mineral soil treatment to enable more realistic modeling of permafrost, an urban canyon model to contrast rural and urban energy balance and climate (CLMU), and an updated biogenic volatile organic compounds (BVOC) model. Other modifications of note include refinement of the global PFT, wetland, and lake distributions, more realistic optical properties for grasslands and croplands, and an improved diurnal cycle and spectral distribution of incoming solar radiation to force the model in land-only mode.

Many of the ideas incorporated into the carbon and nitrogen cycle component of CLM4 derive from the earlier development of the land-only ecosystem process model Biome-BGC (Biome BioGeochemical Cycles), originating at the Numerical Terradynamic Simulation Group (NTSG) at the University of Montana, under the guidance of Prof. Steven Running. Biome-BGC itself is an extension of an earlier model, Forest-BGC ([Running and Coughlan, 1988; Running and Gower, 1991](#)), which simulates water, carbon, and, to a limited extent, nitrogen fluxes for forest ecosystems. Forest-BGC was designed to be driven by remote sensing inputs of vegetation structure, and so used a diagnostic (prescribed) leaf area index, or, in the case of the dynamic allocation version of the model ([Running and Gower, 1991](#)), prescribed maximum leaf area index.

Biome-BGC expanded on the Forest-BGC logic by introducing a more mechanistic calculation of leaf and canopy scale photosynthesis ([Hunt and Running, 1992](#)), and extending the physiological parameterizations to include multiple woody and non-woody vegetation types ([Hunt et al. 1996; Running and Hunt, 1993](#)). Later versions of Biome-BGC introduced more mechanistic descriptions of belowground carbon and nitrogen cycles, nitrogen controls on photosynthesis and decomposition, sunlit and shaded canopies, vertical gradient in leaf morphology, and explicit treatment of fire and harvest disturbance and regrowth dynamics ([Kimball et al. 1997; Thornton, 1998; Thornton et al. 2002; White et al. 2000](#)). Biome-BGC version 4.1.2 ([Thornton et al. 2002](#)) provided a point of departure for integrating new biogeochemistry components into CLM4.

CLM4 was released to the community in June, 2010 along with the Community Climate System Model version 4 (CCSM4). CLM4 is used in CCSM4, CESM1, CESM1.1, and remains available as the default land component model

option for coupled simulations in CESM1.2.

CLM4.5

The motivations for the development of CLM4.5 were similar to those for CLM4: incorporate several recent scientific advances in the understanding and representation of land surface processes, expand model capabilities, and improve surface and atmospheric forcing datasets.

Specifically, several parameterizations were revised to reflect new scientific understanding and in an attempt to reduce biases identified in CLM4 simulations including low soil carbon stocks especially in the Arctic, excessive tropical GPP and unrealistically low Arctic GPP, a dry soil bias in Arctic soils, unrealistically high LAI in the tropics, a transient 20th century carbon response that was inconsistent with observational estimates, and several other more minor problems or biases.

The main modifications include updates to canopy processes including a revised canopy radiation scheme and canopy scaling of leaf processes, co-limitations on photosynthesis, revisions to photosynthetic parameters (*Bonan et al. 2011*), temperature acclimation of photosynthesis, and improved stability of the iterative solution in the photosynthesis and stomatal conductance model (*Sun et al. 2012*). Hydrology updates included modifications such that hydraulic properties of frozen soils are determined by liquid water content only rather than total water content and the introduction of an ice impedance function, and other corrections that increase the consistency between soil water state and water table position and allow for a perched water table above icy permafrost ground (*Swenson et al. 2012*). A new snow cover fraction parameterization is incorporated that reflects the hysteresis in fractional snow cover for a given snow depth between accumulation and melt phases (*Swenson and Lawrence, 2012*). The lake model in CLM4 was replaced with a completely revised and more realistic lake model (*Subin et al. 2012a*). A surface water store was introduced, replacing the wetland land unit and permitting prognostic wetland distribution modeling. The surface energy fluxes are calculated separately (*Swenson and Lawrence, 2012*) for snow-covered, water-covered, and snow/water-free portions of vegetated and crop land units, and snow-covered and snow-free portions of glacier land units. Globally constant river flow velocity is replaced with variable flow velocity based on mean grid cell slope. A vertically resolved soil biogeochemistry scheme is introduced with base decomposition rates modified by soil temperature, water, and oxygen limitations and also including vertical mixing of soil carbon and nitrogen due to bioturbation, cryoturbation, and diffusion (*Koven et al. 2013*). The litter and soil carbon and nitrogen pool structure as well as nitrification and denitrification that were modified based on the Century model. Biological fixation was revised to distribute fixation more realistically over the year (*Koven et al. 2013*). The fire model was replaced with a model that includes representations of natural and anthropogenic triggers and suppression as well as agricultural, deforestation, and peat fires (*Li et al. 2012a,b; Li et al. 2013a*). The biogenic volatile organic compounds model is updated to MEGAN2.1 (*Guenther et al. 2012*).

Additions to the model include a methane production, oxidation, and emissions model (*Riley et al. 2011a*) and an extension of the crop model to include interactive fertilization, organ pools (*Drewniak et al. 2013*), and irrigation (*Sacks et al. 2009*). Elements of the Variable Infiltration Capacity (VIC) model are included as an alternative optional runoff generation scheme (*Li et al. 2011*). There is also an option to run with a multilayer canopy (*Bonan et al. 2012*). Multiple urban density classes, rather than the single dominant urban density class used in CLM4, are modeled in the urban land unit. Carbon (¹³C and ¹⁴C) isotopes are enabled (*Koven et al. 2013*). Minor changes include a switch of the C3 Arctic grass and shrub phenology from stress deciduous to seasonal deciduous and a change in the glacier bare ice albedo to better reflect recent estimates. Finally, the carbon and nitrogen cycle spinup is accelerated and streamlined with a revised spinup method, though the spinup timescale remains long.

Finally, the predominantly low resolution input data provided with CLM4 to create CLM4 surface datasets is replaced with newer and higher resolution input datasets where possible (see section 2.2.3 for details). The default meteorological forcing dataset provided with CLM4 (*Qian et al. 2006*) is replaced with the 1901-2010 CRUNCEP forcing dataset (see Chapter 2.32) for CLM4.5, though users can also still use the *Qian et al. (2006)* dataset or other alternative forcing datasets.

CLM4.5 was released to the community in June 2013 along with the Community Earth System Model version 1.2 (CESM1.2).

CLM5.0

Developments for CLM5.0 build on the progress made in CLM4.5. Most major components of the model have been updated with particularly notable changes made to soil and plant hydrology, snow density, river modeling, carbon and nitrogen cycling and coupling, and crop modeling. Much of the focus of development centered on a push towards more mechanistic treatment of key processes, in addition to more comprehensive and explicit representation of land use and land-cover change. Prior versions of CLM included relatively few options for physics parameterizations or structure. In CLM5, where new parameterizations or model decisions were made, in most cases, the CLM4.5 parameterization was maintained so that users could switch back and forth between different parameterizations via namelist control where appropriate or desirable. Throughout the CLM5 Technical Description, in general only the default parameterization for any given process is described. Readers are referred to the CLM4.5 or CLM4 Technical Descriptions for detailed descriptions of non-default parameterizations.

The hydrology updates include the introduction of a dry surface layer-based soil evaporation resistance parameterization ([Swenson and Lawrence, 2014](#)) and a revised canopy interception parameterization. Canopy interception is now divided into liquid and solid phases, with the intercepted snow subject to unloading events due to wind or above-freezing temperatures. The snow-covered fraction of the canopy is used within the canopy radiation and surface albedo calculation. Instead of applying a spatially uniform soil thickness, soil thickness can vary in space ([Brunke et al. 2016](#) and [Swenson and Lawrence, 2015](#)) and is set to values within a range of 0.4m to 8.5m depth, derived from a spatially explicit soil thickness data product ([Pelletier et al., 2016](#)). The explicit treatment of soil thickness allows for the deprecation of the unconfined aquifer parameterization used in CLM4.5, which is replaced with a zero flux boundary condition and explicit modeling of both the saturated and unsaturated zones. The default model soil layer resolution is increased, especially within the top 3m, to more explicitly represent active layer thickness within the permafrost zone. Rooting profiles were used inconsistently in CLM4.5 with [Zeng \(2001\)](#) profiles used for water and [Jackson et al. \(1996\)](#) profiles used for carbon inputs. For CLM5, the Jackson et al. (1996) rooting profiles are used for both water and carbon. Roots are deepened for the broadleaf evergreen tropical tree and broadleaf deciduous tropical tree types. Finally, an adaptive time-stepping solution to the Richard's equation is introduced, which improves the accuracy and stability of the numerical soil water solution. The River Transport Model (RTM) is replaced with the Model for Scale Adaptive River Transport (MOSART, [Li et al., 2013b](#)) in which surface runoff is routed across hillslopes and then discharged along with subsurface runoff into a tributary subnetwork before entering the main channel.

Several changes are included that are mainly targeted at improving the simulation of surface mass balance over ice sheets. The fresh snow density parameterization is updated to more realistically capture the temperature effects and to additionally account for wind effects on new snow density ([van Kampenhout et al., 2017](#)). The maximum number of snow layers and snow amount is increased from 5 layers and 1m snow water equivalent to 12 layers and 10m snow water equivalent to allow for the formation of firn in regions of persistent snow-cover (e.g., glaciers and ice sheets) ([van Kampenhout et al., 2017](#)). The CISM2 ice sheet model is active for Greenland by default with one-way coupling (surface mass balance impacts ice sheet dynamics, but ice sheet dynamics do not feedback onto surface elevation). Two-way coupling can be activated through a namelist switch. The introduction in CLM5 of the capability to dynamically adjust landunit weights means that a glacier can initiate, grow, shrink, or disappear during a simulation when two-way coupling is active. Multiple elevation classes (10 elevation classes by default) and associated temperature, rain/snow partitioning, and downwelling longwave downscaling are used for glacier landunits to account for the strong topographic elevation heterogeneity over glaciers and ice sheets.

A plant hydraulic stress routine is introduced which explicitly models water transport through the vegetation according to a simple hydraulic framework (Kennedy et al., to be submitted). The water supply equations are used to solve for vegetation water potential forced by transpiration demand and a set of layer-by-layer soil water potentials. Stomatal conductance, therefore, is a function of prognostic leaf water potential. Water stress is calculated as the ratio of attenuated stomatal conductance to maximum stomatal conductance. An emergent feature of the plant hydraulics is soil hydraulic redistribution. In CLM5, maximum stomatal conductance is obtained from the Medlyn conductance model ([Medlyn et al., 2011](#)), rather than the Ball-Berry stomatal conductance model that was utilized in CLM4.5 and prior versions of the model. The Medlyn stomatal conductance model is preferred mainly for its more realistic behavior at low humidity levels ([Rogers et al., 2017](#)). The stress deciduous vegetation phenology trigger is augmented with an antecedent precipitation requirement ([Dahlin et al. 2015](#)).

Plant nutrient dynamics are substantially updated to resolve several deficiencies with the CLM4 and CLM4.5 nutrient

cycling representation. The Fixation and Update of Nitrogen (FUN) model based on the work of [Fisher et al. \(2010\)](#), [Brzostek et al. \(2014\)](#), and [Shi et al. \(2016\)](#) is incorporated. The concept of FUN is that in most cases, N uptake requires the expenditure of energy in the form of carbon, and further, that there are numerous potential sources of N in the environment which a plant may exchange for carbon. The ratio of carbon expended to N acquired is therefore the cost, or exchange rate, of N acquisition. FUN calculates the rate of symbiotic N fixation, with this N passed straight to the plant, not the mineral N pool. Separately, CLM5 also calculates rates of symbiotic (or free living) N fixation as a function of evapotranspiration ([Cleveland et al. 1999](#)), which is added to the soil inorganic ammonium (NH_4^+) pool. The static plant carbon:nitrogen (C:N) ratios utilized in CLM4 and CLM4.5 are replaced with variable plant C:N ratios which allows plants to adjust their C:N ratio, and therefore their leaf nitrogen content, with the cost of N uptake ([Ghimire et al. 2016](#)). The implementation of a flexible C:N ratio means that the model no longer relies on instantaneous downregulation of potential photosynthesis rates based on soil mineral nitrogen availability to represent nutrient limitation. Furthermore, stomatal conductance is now based on the N-limited photosynthesis rather than on potential photosynthesis. Finally, the Leaf Use of Nitrogen for Assimilation (LUNA, [Xu et al., 2012](#) and [Ali et al., 2016](#)) model is incorporated. The LUNA model calculates photosynthetic capacity based on optimization of the use of leaf nitrogen under different environmental conditions such that light capture, carboxylation, and respiration are co-limiting.

CLM5 applies a fixed allocation scheme for woody vegetation. The decision to use a fixed allocation scheme in CLM5, rather than a dynamic NPP-based allocation scheme, as was used in CLM4 and CLM4.5, was driven by the fact that observations indicate that biomass saturates with increasing productivity, in contrast to the behavior in CLM4 and CLM4.5 where biomass continuously increases with increasing productivity ([Negron-Juarez et al., 2015](#)). Soil carbon decomposition processes are unchanged in CLM5, but a new metric for apparent soil carbon turnover times ([Koven et al., 2017](#)) suggested parameter changes that produce a weak intrinsic depth limitation on soil carbon turnover rates (rather than the strong depth limitaiton in CLM4.5) and that the thresholds for soil moisture limitation on soil carbon turnover rates in dry soils should be set at a wetter soil moisture level than that used in CLM4.5.

Representation of human management of the land (agriculture, wood harvest) is augmented in several ways. The CLM4.5 crop model is extended to operate globally through the addition of rice and sugarcane as well as tropical varieties of corn and soybean ([Badger and Dirmeyer, 2015](#) and [Levis et al., 2016](#)). These crop types are added to the existing temperate corn, temperature soybean, spring wheat, and cotton crop types. Fertilization rates and irrigation equipped area updated annually based on crop type and geographic region through an input dataset. The irrigation trigger is updated. Additional minor changes include crop phenological triggers that vary by latitude for selected crop types, grain C and N is now removed at harvest to a 1-year product pool with the carbon for the next season's crop seed removed from the grain carbon at harvest. Through the introduction of the capability to dynamically adjust landunit weights during a simulation, the crop model can now be run coincidentally with prescribed land use, which significantly expands the capabilities of the model. Mass-based rather than area-based wood harvest is applied. Several heat stress indices for both urban and rural areas are calculated and output by default ([Buzan et al., 2015](#)). A more sophisticated and realistic building space heating and air conditioning submodel that prognoses interior building air temperature and includes more realistic space heating and air conditioning wasteheat factors is incorporated.

The fire model is the same as utilized in CLM4.5 except that a modified scheme is used to estimate the dependence of fire occurrence and spread on fuel wetness for non-peat fires outside cropland and tropical closed forests ([Li and Lawrence, 2017](#)) and the dependence of agricultural fires on fuel load is removed.

Included with the release of CLM5.0 is a functionally supported version of the Functionally-Assembled Terrestrial Ecosystem Simulator (FATES, [Fisher et al., 2015](#)). A major motivation of FATES is to allow the prediction of biome boundaries directly from plant physiological traits via their competitive interactions. FATES is a cohort model of vegetation competition and co-existence, allowing a representation of the biosphere which accounts for the division of the land surface into successional stages, and for competition for light between height structured cohorts of representative trees of various plant functional types. FATES is not active by default in CLM5.0.

Note that the classical dynamic global vegetation model (CLM-DGVM) that has been available within CLM4 and CLM4.5 remains available, though it is largely untested. The technical description of the CLM-DGVM can be found within the CLM4.5 Technical Description ([Oleson et al. 2013](#)).

During the course of the development of CLM5.0, it became clear that the increasing complexity of the model combined with the increasing number and range of model development projects required updates to the underlying CLM

infrastructure. Many such software improvements are included in CLM5 including a partial transition to an object-oriented modular software structure. Many hard coded model parameters have been extracted into either the parameter file or the CLM namelist, which allows users to more readily calibrate the model for use at specific locations or to conduct parameter sensitivity studies. As part of the effort to increase the scientific utility of the code, in most instances older generation parameterizations (i.e., the parameterizations available in CLM4 or CLM4.5) are retained under namelist switches, allowing the user to revert to CLM4.5 from the same code base or to revert individual parameterizations where the old parameterizations are compatible with the new code. Finally, multiple vertical soil layer structures are defined and it is relatively easy to add additional structures.

2.1.2 Biogeophysical and Biogeochemical Processes

Biogeophysical and biogeochemical processes are simulated for each subgrid land unit, column, and plant functional type (PFT) independently and each subgrid unit maintains its own prognostic variables (see section 2.2.1 for definitions of subgrid units). The same atmospheric forcing is used to force all subgrid units within a grid cell. The surface variables and fluxes required by the atmosphere are obtained by averaging the subgrid quantities weighted by their fractional areas. The processes simulated include (Figure 2.1):

1. Surface characterization including land type heterogeneity and ecosystem structure (Chapter 2.2)
2. Absorption, reflection, and transmittance of solar radiation (Chapter 2.3, 2.4)
3. Absorption and emission of longwave radiation (Chapter 2.4)
4. Momentum, sensible heat (ground and canopy), and latent heat (ground evaporation, canopy evaporation, transpiration) fluxes (Chapter 2.5)
5. Heat transfer in soil and snow including phase change (Chapter 2.6)
6. Canopy hydrology (interception, throughfall, and drip) (Chapter 2.7)
7. Soil hydrology (surface runoff, infiltration, redistribution of water within the column, sub-surface drainage, groundwater) (Chapter 2.7)
8. Snow hydrology (snow accumulation and melt, compaction, water transfer between snow layers) (Chapter 2.8)
9. Stomatal physiology, photosynthetic capacity, and photosynthesis (Chapters 2.9 and 2.10)
10. Plant hydraulics (Chapter 2.11)
11. Lake temperatures and fluxes (Chapter 2.12)
12. Glacier processes (Chapter 2.13)
13. River routing and river flow (Chapter 2.14)
14. Urban energy balance and climate (Chapter 2.15)
15. Vegetation carbon and nitrogen allocation (Chapter 2.19)
16. Vegetation phenology (Chapter 2.20)
17. Plant respiration (Chapter 2.17)
18. Soil and litter carbon decomposition (Chapter 2.21)
19. Fixation and uptake of nitrogen (Chapter 2.18)
20. External nitrogen cycling including deposition, denitrification, leaching, and losses due to fire (Chapter 2.22)
21. Plant mortality (Chapter `rst_Plant_Mortality`)
22. Fire ignition, suppression, spread, and emissions, including natural, deforestation, and agricultural fire (Chapter 2.23)
23. Methane production, oxidation, and emissions (Chapter 2.24)

24. Crop dynamics, irrigation, and fertilization (Chapter 2.25)
25. Land cover and land use change including wood harvest (Chapter 2.26)
26. Biogenic volatile organic compound emissions (Chapter 2.29)
27. Dust mobilization and deposition (Chapter 2.30)
28. Carbon isotope fractionation (Chapter 2.31)

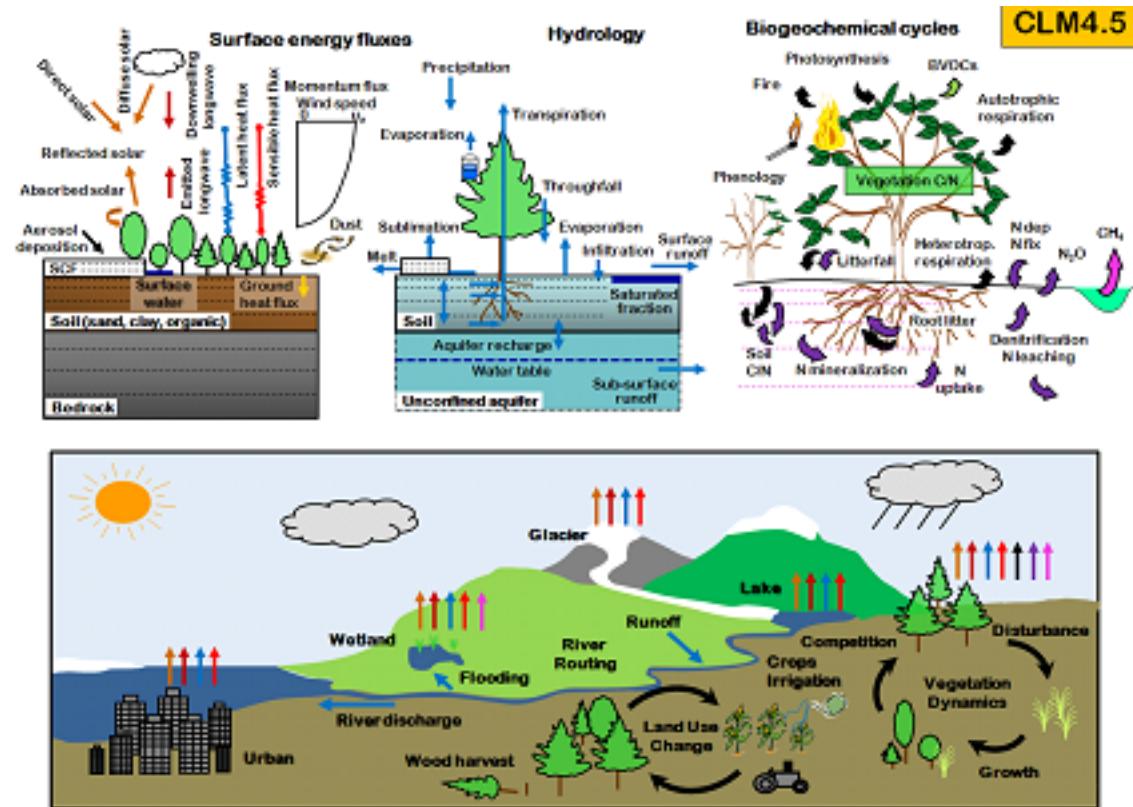


Figure 2.1: Land biogeophysical, biogeochemical, and landscape processes simulated by CLM (adapted from [Lawrence et al. \(2011\)](#) for CLM4.5).

2.2 Surface Characterization, Vertical Discretization, and Model Input Requirements

2.2.1 Surface Characterization

Surface Heterogeneity and Data Structure

Spatial land surface heterogeneity in CLM is represented as a nested subgrid hierarchy in which grid cells are composed of multiple land units, snow/soil columns, and PFTs (Figure 2.2). Each grid cell can have a different number of land units, each land unit can have a different number of columns, and each column can have multiple PFTs. The first subgrid level, the land unit, is intended to capture the broadest spatial patterns of subgrid heterogeneity. The current land units are glacier, lake, urban, vegetated, and crop (when the crop model option is turned on). The land unit

level can be used to further delineate these patterns. For example, the urban land unit is divided into density classes representing the tall building district, high density, and medium density urban areas.

The second subgrid level, the column, is intended to capture potential variability in the soil and snow state variables within a single land unit. For example, the vegetated land unit could contain several columns with independently evolving vertical profiles of soil water and temperature. Similarly, the managed vegetation land unit can be divided into two columns, irrigated and non-irrigated. The default snow/soil column is represented by 25 layers for ground (with up to 20 of these layers classified as soil layers and the remaining layers classified as bedrock layers) and up to 10 layers for snow, depending on snow depth. The central characteristic of the column subgrid level is that this is where the state variables for water and energy in the soil and snow are defined, as well as the fluxes of these components within the soil and snow. Regardless of the number and type of PFTs occupying space on the column, the column physics operates with a single set of upper boundary fluxes, as well as a single set of transpiration fluxes from multiple soil levels. These boundary fluxes are weighted averages over all PFTs. Currently, for glacier, lake, and vegetated land units, a single column is assigned to each land unit. The crop land unit is split into irrigated and unirrigated columns with a single crop occupying each column. The urban land units have five columns (roof, sunlit walls and shaded walls, and pervious and impervious canyon floor) (Oleson et al. 2010b).

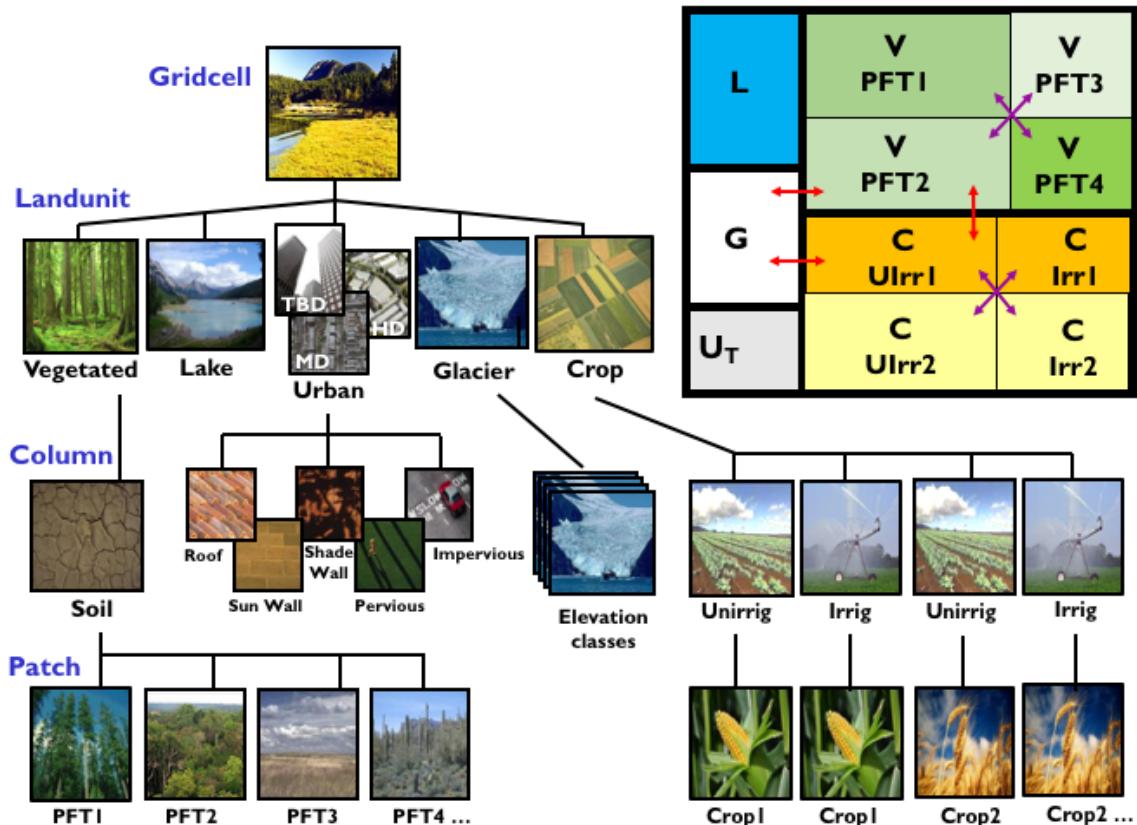


Figure 2.2: Configuration of the CLM subgrid hierarchy. Box in upper right shows hypothetical subgrid distribution for a single grid cell. Note that the Crop land unit is only used when the model is run with the crop model active. Abbreviations: TBD – Tall Building District; HD – High Density; MD – Medium Density, G – Glacier, L – Lake, U – Urban, C – Crop, V – Vegetated, PFT – Plant Functional Type, Irr – Irrigated, UIrr – Unirrigated. Red arrows indicate allowed land unit transitions. Purple arrows indicate allowed patch-level transitions.

The third subgrid level is referred to as the patch level. Patches can be PFTs or bare ground on the vegetated land

unit and crop functional types (CFTs) on the crop land unit. The patch level is intended to capture the biogeophysical and biogeochemical differences between broad categories of plants in terms of their functional characteristics. On the vegetated land unit, up to 16 possible PFTs that differ in physiology and structure may coexist on a single column. All fluxes to and from the surface are defined at the PFT level, as are the vegetation state variables (e.g. vegetation temperature and canopy water storage). On the crop land unit, typically, different crop types can be represented on each

crop land unit column (see Chapter 2.25 for details).

In addition to state and flux variable data structures for conserved components at each subgrid level (e.g., energy, water, carbon), each subgrid level also has a physical state data structure for handling quantities that are not involved in conservation checks (diagnostic variables). For example, the urban canopy air temperature and humidity are defined through physical state variables at the land unit level, the number of snow layers and the soil roughness lengths are defined as physical state variables at the column level, and the leaf area index and the fraction of canopy that is wet are defined as physical state variables at the PFT level.

The standard configuration of the model subgrid hierarchy is illustrated in Figure 2.2. Here, only four PFTs are shown associated with the single column beneath the vegetated land unit but up to sixteen are possible. The crop land unit is present only when the crop model is active.

Note that the biogeophysical processes related to soil and snow require PFT level properties to be aggregated to the column level. For example, the net heat flux into the ground is required as a boundary condition for the solution of snow/soil temperatures (Chapter 2.6). This column level property must be determined by aggregating the net heat flux from all PFTs sharing the column. This is generally accomplished in the model by computing a weighted sum of the desired quantity over all PFTs whose weighting depends on the PFT area relative to all PFTs, unless otherwise noted in the text.

Vegetation Composition

Vegetated surfaces are comprised of up to 15 possible plant functional types (PFTs) plus bare ground (Table 2.1). An additional PFT is added if the irrigation model is active and six additional PFTs are added if the crop model is active (Chapter 2.25). These plant types differ in leaf and stem optical properties that determine reflection, transmittance, and absorption of solar radiation (Table 2.8), root distribution parameters that control the uptake of water from the soil (Table 2.23), aerodynamic parameters that determine resistance to heat, moisture, and momentum transfer (Table 2.16), and photosynthetic parameters that determine stomatal resistance, photosynthesis, and transpiration (Table 2.21, Table 2.22). The composition and abundance of PFTs within a grid cell can either be prescribed as time-invariant fields (e.g., using the present day dataset described in section 21.3.3) or can evolve with time if the model is run in transient landcover mode (Chapter 2.26).

Table 2.1: Plant functional types

Plant functional type	Acronym
Needleleaf evergreen tree – temperate	NET Temperate
Needleleaf evergreen tree - boreal	NET Boreal
Needleleaf deciduous tree – boreal	NDT Boreal
Broadleaf evergreen tree – tropical	BET Tropical
Broadleaf evergreen tree – temperate	BET Temperate
Broadleaf deciduous tree – tropical	BDT Tropical
Broadleaf deciduous tree – temperate	BDT Temperate
Broadleaf deciduous tree – boreal	BDT Boreal
Broadleaf evergreen shrub - temperate	BES Temperate
Broadleaf deciduous shrub – temperate	BDS Temperate
Broadleaf deciduous shrub – boreal	BDS Boreal
C ₃ arctic grass	•
C ₃ grass	•
C ₄ grass	•
C ₃ Unmanaged Rainfed Crop	UCrop UIrr
¹ C ₃ Unmanaged Irrigated Crop	UCrop Irr
² Managed Rainfed Unirrigated Crops	Crop UIrr
² Managed Irrigated Crops	Crop Irr

¹Only used if irrigation is active (Chapter 2.25). ²Only used if crop model is active (see Chapter 2.25 for list of represented crops).

Vegetation Structure

Vegetation structure is defined by leaf and stem area indices (L , S) and canopy top and bottom heights (z_{top}, z_{bot}). Separate leaf and stem area indices and canopy heights are prescribed or calculated for each PFT. Daily leaf and stem area indices are obtained from gridded datasets of monthly values (section 2.2.3). Canopy top and bottom heights for trees are from ICESat (*Simard et al. (2011)*). Canopy top and bottom heights for short vegetation are obtained from gridded datasets but are invariant in space and time and were obtained from PFT-specific values (*Bonan et al. (2002a)*) (Table 2.2). When the biogeochemistry model is active, vegetation state (LAI, SAI, canopy top and bottom heights) are calculated prognostically (see Chapter 2.20).

Table 2.2: Plant functional type canopy top and bottom heights

Plant functional type	z_{top}	z_{bot}
BES Temperate	0.5	0.1
BDS Temperate	0.5	0.1
BDS Boreal	0.5	0.1
C ₃ arctic grass	0.5	0.01
C ₃ grass	0.5	0.01
C ₄ grass	0.5	0.01
UCrop UIrr	0.5	0.01
UCrop Irr	0.5	0.01
Crop UIrr	0.5	0.01
Crop Irr	0.5	0.01

Phenology and vegetation burial by snow

When the biogeochemistry model is inactive, leaf and stem area indices ($\text{m}^2 \text{ leaf area m}^{-2}$ ground area) are updated daily by linearly interpolating between monthly values. Monthly PFT leaf area index values are developed from the 1-km MODIS-derived monthly grid cell average leaf area index of *Myndeni et al. (2002)*, as described in *Lawrence and Chase (2007)*. Stem area index is calculated from the monthly PFT leaf area index using the methods of *Zeng et al. (2002)*. The leaf and stem area indices are adjusted for vertical burying by snow (*Wang and Zeng 2009*) as

$$A = A^*(1 - f_{veg}^{sno}) \quad (2.1)$$

where A is the leaf or stem area before adjustment for snow, A^* is the remaining exposed leaf or stem area, f_{veg}^{sno} is the vertical fraction of vegetation covered by snow

$$\begin{aligned} f_{veg}^{sno} &= \frac{z_{sno} - z_{bot}}{z_{top} - z_{bot}} && \text{for tree and shrub} \\ f_{veg}^{sno} &= \frac{\min(z_{sno}, z_c)}{z_c} && \text{for grass and crop} \end{aligned} \quad (2.2)$$

where $z_{sno} - z_{bot} \geq 0$, $0 \leq f_{veg}^{sno} \leq 1$, z_{sno} is the depth of snow (m) (Chapter 2.8), and $z_c = 0.2$ is the snow depth when short vegetation is assumed to be completely buried by snow (m). For numerical reasons, exposed leaf and stem area are set to zero if less than 0.05. If the sum of exposed leaf and stem area is zero, then the surface is treated as snow-covered ground.

2.2.2 Vertical Discretization

Soil Layers

The soil column can be discretized into an arbitrary number of layers. The default vertical discretization (Table 2.3) uses $N_{levgrnd} = 25$ layers, of which $N_{levsoi} = 20$ are hydrologically and biogeochemically active. The deepest 5 layers are only included in the thermodynamical calculations (*Lawrence et al. 2008*) described in Chapter 2.6.

The layer structure of the soil is described by the node depth, z_i (m), the thickness of each layer, Δz_i (m), and the depths at the layer interfaces $z_{h,i}$ (m).

Table 2.3: Soil layer structure

Layer	z_i	Δz_i	$z_{h,i}$
1	0.010	0.020	0.020
2	0.040	0.040	0.060
3	0.090	0.060	0.120
4	0.160	0.080	0.200
5	0.260	0.120	0.320
6	0.400	0.160	0.480
7	0.580	0.200	0.680
8	0.800	0.240	0.920
9	1.060	0.280	1.200
10	1.360	0.320	1.520
11	1.700	0.360	1.880
12	2.080	0.400	2.280
13	2.500	0.440	2.720
14	2.990	0.540	3.260
15	3.580	0.640	3.900
16	4.270	0.740	4.640
17	5.060	0.840	5.480
18	5.950	0.940	6.420
19	6.940	1.040	7.460
20	8.030	1.140	8.600
21	9.795	2.390	10.990
22	13.328	4.676	15.666
23	19.483	7.635	23.301
24	28.871	11.140	34.441
25	41.998	15.115	49.556

Layer node depth (z_i), thickness (Δz_i), and depth at layer interface ($z_{h,i}$) for default soil column. All in meters.

Depth to Bedrock

The hydrologically and biogeochemically active portion of the soil column can be restricted to a thickness less than that of the maximum soil depth. By providing a depth-to-bedrock dataset, which may vary spatially, the number of layers used in the hydrologic and biogeochemical calculations, $N_{bedrock}$, may be specified, subject to the constraint $N_{bedrock} \leq N_{levsoi}$

2.2.3 Model Input Requirements

Atmospheric Coupling

The current state of the atmosphere (Table 2.4) at a given time step is used to force the land model. This atmospheric state is provided by an atmospheric model in coupled mode or from an observed dataset in land-only mode (Chapter 2.32). The land model then initiates a full set of calculations for surface energy, constituent, momentum, and radiative fluxes. The land model calculations are implemented in two steps. The land model proceeds with the calculation of surface energy, constituent, momentum, and radiative fluxes using the snow and soil hydrologic states from the previous time step. The land model then updates the soil and snow hydrology calculations based on these fluxes. These fields are passed to the atmosphere (Table 2.5). The albedos sent to the atmosphere are for the solar zenith angle at the next time step but with surface conditions from the current time step.

Table 2.4: Atmospheric input to land model

Field	variable name	units
¹ Reference height	z'_{atm}	m
Atmosphere model's surface height	$z_{surf,atm}$	m
Zonal wind at z_{atm}	u_{atm}	m s^{-1}
Meridional wind at z_{atm}	v_{atm}	m s^{-1}
Potential temperature	θ_{atm}	K
Specific humidity at z_{atm}	q_{atm}	kg kg^{-1}
Pressure at z_{atm}	P_{atm}	Pa
Temperature at z_{atm}	T_{atm}	K
Incident longwave radiation	$L_{atm} \downarrow$	W m^{-2}
² Liquid precipitation	q_{rain}	mm s^{-1}
² Solid precipitation	q_{sno}	mm s^{-1}
Incident direct beam visible solar radiation	$S_{atm} \downarrow_{vis}^{\mu}$	W m^{-2}
Incident direct beam near-infrared solar radiation	$S_{atm} \downarrow_{nir}^{\mu}$	W m^{-2}
Incident diffuse visible solar radiation	$S_{atm} \downarrow_{vis}$	W m^{-2}
Incident diffuse near-infrared solar radiation	$S_{atm} \downarrow_{nir}$	W m^{-2}
Carbon dioxide (CO_2) concentration	c_a	ppmv
³ Aerosol deposition rate	D_{sp}	$\text{kg m}^{-2} \text{s}^{-1}$
⁴ Nitrogen deposition rate	NF_{ndep_sminn}	$\text{g (N) m}^{-2} \text{yr}^{-1}$
⁵ Lightning frequency	I_l	flash $\text{km}^{-2} \text{hr}^{-1}$

¹The atmospheric reference height received from the atmospheric model z'_{atm} is assumed to be the height above the surface as defined by the roughness length z_0 plus displacement height d . Thus, the reference height used for flux computations (Chapter 2.5) is $z_{atm} = z'_{atm} + z_0 + d$. The reference heights for temperature, wind, and specific humidity ($z_{atm,h}$, $z_{atm,m}$, $z_{atm,w}$) are required. These are set equal to z_{atm} .

²CAM provides convective and large-scale liquid and solid precipitation, which are added to yield total liquid precipitation q_{rain} and solid precipitation q_{sno} .

³There are 14 aerosol deposition rates required depending on species and affinity for bonding with water; 8 of these are dust deposition rates (dry and wet rates for 4 dust size bins, $D_{dst,dry1}$, $D_{dst,dry2}$, $D_{dst,dry3}$, $D_{dst,dry4}$, $D_{dst,wet1}$, $D_{dst,wet2}$, $D_{dst,wet3}$, $D_{dst,wet4}$), 3 are black carbon deposition rates (dry and wet hydrophilic and dry hydrophobic rates, $D_{bc,dryphil}$, $D_{bc,wetphil}$, $D_{bc,dryphob}$), and 3 are organic carbon deposition rates (dry and wet hydrophilic and dry hydrophobic rates, $D_{oc,dryphil}$, $D_{oc,wetphil}$, $D_{oc,dryphob}$). These fluxes are computed interactively by the atmospheric model (when prognostic aerosol representation is active) or are prescribed from a time-varying (annual cycle or transient), globally-gridded deposition file defined in the namelist (see the CLM4.5 User's Guide). Aerosol deposition rates were calculated in a transient 1850-2009 CAM simulation (at a resolution of 1.9x2.5x26L) with interactive chemistry (troposphere and stratosphere) driven by CCSM3 20th century sea-surface temperatures and emissions (Lamarque et al. 2010) for short-lived gases and aerosols; observed concentrations were specified for methane, N_2O , the ozone-depleting substances (CFCs), and CO_2 . The fluxes are used by the snow-related parameterizations (Chapters 2.3 and numref:rst_Snow_Hydrology).

⁴The nitrogen deposition rate is required by the biogeochemistry model when active and represents the total deposition of mineral nitrogen onto the land surface, combining deposition of NO_y and NH_x . The rate is supplied either as a time-invariant spatially-varying annual mean rate or time-varying for a transient simulation. Nitrogen deposition rates were calculated from the same CAM chemistry simulation that generated the aerosol deposition rates.

⁵Climatological 3-hourly lightning frequency at $\sim 1.8^\circ$ resolution is provided, which was calculated via bilinear interpolation from 1995-2011 NASA LIS/OTD grid product v2.2 (<http://ghrc.msfc.nasa.gov>) 2-hourly, 2.5° lightning frequency data. In future versions of the model, lightning data may be obtained directly from the atmosphere model.

Density of air (ρ_{atm}) (kg m^{-3}) is also required but is calculated directly from $\rho_{atm} = \frac{P_{atm} - 0.378e_{atm}}{R_{da}T_{atm}}$ where P_{atm} is atmospheric pressure (Pa), e_{atm} is atmospheric vapor pressure (Pa), R_{da} is the gas constant for dry air (J kg^{-1}

K^{-1}) (Table 2.7), and T_{atm} is the atmospheric temperature (K). The atmospheric vapor pressure e_{atm} is derived from atmospheric specific humidity q_{atm} (kg kg^{-1}) as $e_{atm} = \frac{q_{atm} P_{atm}}{0.622 + 0.378 q_{atm}}$.

The O_2 partial pressure (Pa) is required but is calculated from molar ratio and the atmospheric pressure P_{atm} as $o_i = 0.209 P_{atm}$.

Table 2.5: Land model output to atmospheric model

Field	Variable name	units
¹ Latent heat flux	$\lambda_{vap} E_v + \lambda E_g$	W m^{-2}
Sensible heat flux	$H_v + H_g$	W m^{-2}
Water vapor flux	$E_v + E_g$	mm s^{-1}
Zonal momentum flux	τ_x	$\text{kg m}^{-1} \text{s}^{-2}$
Meridional momentum flux	τ_y	$\text{kg m}^{-1} \text{s}^{-2}$
Emitted longwave radiation	$L \uparrow$	W m^{-2}
Direct beam visible albedo	$I \uparrow_{vis}^\mu$	•
Direct beam near-infrared albedo	$I \uparrow_{nir}^\mu$	•
Diffuse visible albedo	$I \uparrow_{vis}$	•
Diffuse near-infrared albedo	$I \uparrow_{nir}$	•
Absorbed solar radiation	\vec{S}	W m^{-2}
Radiative temperature	T_{rad}	K
Temperature at 2 meter height	T_{2m}	K
Specific humidity at 2 meter height	q_{2m}	kg kg^{-1}
Wind speed at 10 meter height	u_{10m}	m s^{-1}
Snow water equivalent	W_{sno}	m
Aerodynamic resistance	r_{am}	s m^{-1}
Friction velocity	u_*	m s^{-1}
² Dust flux	F_j	$\text{kg m}^{-2} \text{s}^{-1}$
Net ecosystem exchange	NEE	$\text{kg CO}_2 \text{ m}^{-2} \text{s}^{-1}$

¹ λ_{vap} is the latent heat of vaporization (J kg^{-1}) (Table 2.7) and λ is either the latent heat of vaporization λ_{vap} or latent heat of sublimation λ_{sub} (J kg^{-1}) (Table 2.7) depending on the liquid water and ice content of the top snow/soil layer (section 5.4).

²There are $j = 1, \dots, 4$ dust transport bins.

Initialization

Initialization of the land model (i.e., providing the model with initial temperature and moisture states) depends on the type of run (startup or restart) (see the CLM4.5 User's Guide). A startup run starts the model from either initial conditions that are set internally in the Fortran code (referred to as arbitrary initial conditions) or from an initial conditions dataset that enables the model to start from a spun up state (i.e., where the land is in equilibrium with the simulated climate). In restart runs, the model is continued from a previous simulation and initialized from a restart file that ensures that the output is bit-for-bit the same as if the previous simulation had not stopped. The fields that are required from the restart or initial conditions files can be obtained by examining the code. Arbitrary initial conditions are specified as follows.

Soil points are initialized with surface ground temperature T_g and soil layer temperature T_i , for $i = 1, \dots, N_{levgrnd}$, of 274 K, vegetation temperature T_v of 283 K, no snow or canopy water ($W_{sno} = 0$, $W_{can} = 0$), and volumetric

soil water content $\theta_i = 0.15 \text{ mm}^3 \text{ mm}^{-3}$ for layers $i = 1, \dots, N_{levsoi}$ and $\theta_i = 0.0 \text{ mm}^3 \text{ mm}^{-3}$ for layers $i = N_{levsoi} + 1, \dots, N_{levgrnd}$. placeLake temperatures (T_g and T_i) are initialized at 277 K and $W_{sno} = 0$.

Glacier temperatures ($T_g = T_{snl+1}$ and T_i for $i = snl + 1, \dots, N_{levgrnd}$ where snl is the negative of the number of snow layers, i.e., snl ranges from -5 to 0) are initialized to 250 K with a snow water equivalent $W_{sno} = 1000 \text{ mm}$, snow depth $z_{sno} = \frac{W_{sno}}{\rho_{sno}}$ (m) where $\rho_{sno} = 250 \text{ kg m}^{-3}$ is an initial estimate for the bulk density of snow, and $\theta_i = 1.0$ for $i = 1, \dots, N_{levgrnd}$. The snow layer structure (e.g., number of snow layers snl and layer thickness) is initialized based on the snow depth (section 6.1). The snow liquid water and ice contents (kg m^{-2}) are initialized as $w_{liq,i} = 0$ and $w_{ice,i} = \Delta z_i \rho_{sno}$, respectively, where $i = snl + 1, \dots, 0$ are the snow layers, and Δz_i is the thickness of snow layer i (m). The soil liquid water and ice contents are initialized as $w_{liq,i} = 0$ and $w_{ice,i} = \Delta z_i \rho_{ice} \theta_i$ for $T_i \leq T_f$, and $w_{liq,i} = \Delta z_i \rho_{liq} \theta_i$ and $w_{ice,i} = 0$ for $T_i > T_f$, where ρ_{ice} and ρ_{liq} are the densities of ice and liquid water (kg m^{-3}) (Table 2.7), and T_f is the freezing temperature of water (K) (Table 2.7). All vegetated and glacier land units are initialized with water stored in the unconfined aquifer and unsaturated soil $W_a = 4000 \text{ mm}$ and water table depth z_∇ at five meters below the soil column.

Surface Data

Required surface data for each land grid cell are listed in Table 2.6 and include the glacier, lake, and urban fractions of the grid cell (vegetated and crop occupy the remainder), the fractional cover of each plant functional type (PFT), monthly leaf and stem area index and canopy top and bottom heights for each PFT, soil color, soil texture, soil organic matter density, maximum fractional saturated area, slope, elevation, biogenic volatile organic compounds (BVOCs) emissions factors, population density, gross domestic production, peat area fraction, and peak month of agricultural burning. Optional surface data include crop irrigation and managed crops. All fields are aggregated to the model's grid from high-resolution input datasets (Table 2.6) that are obtained from a variety of sources described below.

Table 2.6: Surface data required for CLM and their base spatial resolution

Surface Field	Resolution
Percent glacier	0.05°
Percent lake and lake depth	0.05°
Percent urban	0.05°
Percent plant functional types (PFTs)	0.05°
Monthly leaf and stem area index	0.5°
Canopy height (top, bottom)	0.5°
Soil color	0.5°
Percent sand, percent clay	0.083°
Soil organic matter density	0.083°
Maximum fractional saturated area	0.125°
Elevation	1km
Slope	1km
Biogenic Volatile Organic Compounds	0.5°
Crop Irrigation	0.083°
Managed crops	0.5°
Population density	0.5°
Gross domestic production	0.5°
Peat area fraction	0.5°
Peak month of agricultural waste burning	0.5°

At the base spatial resolution of 0.05°, the percentage of each PFT is defined with respect to the vegetated portion of the grid cell and the sum of the PFTs is 100%. The percent lake, wetland, glacier, and urban at their base resolution are specified with respect to the entire grid cell. The surface dataset creation routines re-adjust the PFT percentages to ensure that the sum of all land cover types in the grid cell sum to 100%. A minimum threshold of 0.1% of the grid cell by area is required for urban areas.

The percentage glacier mask was derived from vector data of global glacier and ice sheet spatial coverage. Vector data for glaciers (ice caps, icefields and mountain glaciers) were taken from the first globally complete glacier inventory, the Randolph Glacier Inventory version 1.0 (RGIV1.0: [Arendt et al. 2012](#)). Vector data for the Greenland Ice Sheet were provided by Frank Paul and Tobias Bolch (University of Zurich: [Rastner et al. 2012](#)). Antarctic Ice Sheet data were provided by Andrew Bliss (University of Alaska) and were extracted from the Scientific Committee on Antarctic Research (SCAR) Antarctic Digital Database version 5.0. Floating ice is only provided for the Antarctic and does not include the small area of Arctic ice shelves. High spatial resolution vector data were then processed to determine the area of glacier, ice sheet and floating ice within 30-second grid cells globally. The 30-second glacier, ice sheet and Antarctic ice shelf masks were subsequently draped over equivalent-resolution GLOBE topography (Global Land One-km Base Elevation Project, Hastings et al. 1999) to extract approximate ice-covered elevations of ice-covered regions. Grid cells flagged as land-ice in the mask but ocean in GLOBE (typically, around ice sheets at high latitudes) were designated land-ice with an elevation of 0 meters. Finally, the high-resolution mask/topography datasets were aggregated and processed into three 3-minute datasets: 3-minute fractional areal land ice coverage (including both glaciers and ice sheets); 3-minute distributions of areal glacier fractional coverage by elevation and areal ice sheet fractional coverage by elevation. Ice fractions were binned at 100 meter intervals, with bin edges defined from 0 to 6000 meters (plus one top bin encompassing all remaining high-elevation ice, primarily in the Himalaya). These distributions by elevation are needed when running CLM4 with multiple glacier elevation classes.

Percent lake and lake depth are area-averaged from the 90-second resolution data of [Kourzeneva \(2009, 2010\)](#) to the 0.05° resolution using the MODIS land-mask. Percent urban is derived from LandScan 2004, a population density dataset derived from census data, nighttime lights satellite observations, road proximity and slope ([Dobson et al. 2000](#)) as described by [Jackson et al. \(2010\)](#) at 1km resolution and aggregated to 0.05°. A number of urban radiative, thermal, and morphological fields are also required and are obtained from [Jackson et al. \(2010\)](#). Their description can be found in Table 3 of the Community Land Model Urban (CLMU) technical note ([Oleson et al. 2010b](#)).

Percent PFTs are derived from MODIS satellite data as described in [Lawrence and Chase \(2007\)](#) (section 21.3.3). Prescribed PFT leaf area index is derived from the MODIS satellite data of [Myneni et al. \(2002\)](#) using the de-aggregation methods described in [Lawrence and Chase \(2007\)](#) (section 2.2.3). Prescribed PFT stem area index is derived from PFT leaf area index phenology combined with the methods of [Zeng et al. \(2002\)](#). Prescribed canopy top and bottom heights are from [Bonan \(1996\)](#) as described in [Bonan et al. \(2002b\)](#). If the biogeochemistry model is active, it supplies the leaf and stem area index and canopy top and bottom heights dynamically, and the prescribed values are ignored.

Soil color determines dry and saturated soil albedo (section 2.3.2). Soil colors are from [Lawrence and Chase \(2007\)](#).

The soil texture and organic matter content determine soil thermal and hydrologic properties (sections 6.3 and 7.4.1). The International Geosphere-Biosphere Programme (IGBP) soil dataset (Global Soil Data Task 2000) of 4931 soil mapping units and their sand and clay content for each soil layer were used to create a mineral soil texture dataset ([Bonan et al. 2002b](#)). Soil organic matter data is merged from two sources. The majority of the globe is from ISRIC-WISE ([Batjes, 2006](#)). The high latitudes come from the 0.25° version of the Northern Circumpolar Soil Carbon Database ([Hugelius et al. 2012](#)). Both datasets report carbon down to 1m depth. Carbon is partitioned across the top seven CLM4 layers (~1m depth) as in [Lawrence and Slater \(2008\)](#).

The maximum fractional saturated area (f_{\max}) is used in determining surface runoff and infiltration (section 7.3). Maximum fractional saturated area at 0.125° resolution is calculated from 1-km compound topographic indices (CTIs) based on the USGS HYDRO1K dataset ([Verdin and Greenlee 1996](#)) following the algorithm in [Niu et al. \(2005\)](#). f_{\max} is the ratio between the number of 1-km pixels with CTIs equal to or larger than the mean CTI and the total number of pixels in a 0.125° grid cell. See section 7.3.1 and [Li et al. \(2013b\)](#) for further details. Slope and elevation are also obtained from the USGS HYDRO1K 1-km dataset ([Verdin and Greenlee 1996](#)). Slope is used in the surface water parameterization (section 2.7.2), and elevation is used to calculate the grid cell standard deviation of topography for the snow cover fraction parameterization (section 2.8.1).

Biogenic Volatile Organic Compounds emissions factors are from the Model of Emissions of Gases and Aerosols from Nature version 2.1 (MEGAN2.1; [Guenther et al. 2012](#)).

The default list of PFTs includes an unmanaged crop treated as a second C3 grass ([Table 2.1](#)). The unmanaged crop has grid cell fractional cover assigned from MODIS satellite data ([Lawrence and Chase \(2007\)](#)). A managed crop option

uses grid cell fractional cover from the present-day crop dataset of [Ramankutty and Foley \(1998\)](#) (CLM4CNcrop). Managed crops are assigned in the proportions given by [Ramankutty and Foley \(1998\)](#) without exceeding the area previously assigned to the unmanaged crop. The unmanaged crop continues to occupy any of its original area that remains and continues to be handled just by the CN part of CLM4CNcrop. The managed crop types (corn, soybean, and temperate cereals) were chosen based on the availability of corresponding algorithms in AgroIBIS ([Kucharik et al. 2000; Kucharik and Brye 2003](#)). Temperate cereals include wheat, barley, and rye here. All temperate cereals are treated as summer crops (like spring wheat, for example) at this time. Winter cereals (such as winter wheat) may be introduced in a future version of the model.

To allow crops to coexist with natural vegetation in a grid cell and be treated by separate models (i.e., CLM4.5BGCcrop versus the Dynamic Vegetation version (CLM4.5BGCDV)), we separate the vegetated land unit into a naturally vegetated land unit and a human managed land unit. PFTs in the naturally vegetated land unit share one soil column and compete for water (default CLM setting). PFTs in the human managed land unit do not share soil columns and thus permit for differences in land management between crops.

CLM includes the option to irrigate cropland areas that are equipped for irrigation. The application of irrigation responds dynamically to climate (see Chapter 2.25). In CLM, irrigation is implemented for the C3 generic crop only. When irrigation is enabled, the cropland area of each grid cell is divided into an irrigated and unirrigated fraction according to a dataset of areas equipped for irrigation ([Siebert et al. \(2005\)](#)). The area of irrigated cropland in each grid cell is given by the smaller of the grid cell's total cropland area, according to the default CLM4 dataset, and the grid cell's area equipped for irrigation. The remainder of the grid cell's cropland area (if any) is then assigned to unirrigated cropland. Irrigated and unirrigated crops are placed on separate soil columns, so that irrigation is only applied to the soil beneath irrigated crops.

Several input datasets are required for the fire model ([Li et al. 2013a](#)) including population density, gross domestic production, peat area fraction, and peak month of agricultural waste burning. Population density at 0.5° resolution for 1850-2100 combines 5-min resolution decadal population density data for 1850–1980 from the Database of the Global Environment version 3.1 (HYDEv3.1) with 0.5° resolution population density data for 1990, 1995, 2000, and 2005 from the Gridded Population of the World version 3 dataset (GPWv3) (CIESIN, 2005). Gross Domestic Production (GDP) per capita in 2000 at 0.5° is from [Van Vuuren et al. \(2006\)](#), which is the base-year GDP data for IPCC-SRES and derived from country-level World Bank's World Development Indicators (WDI) measured in constant 1995 US\$ ([World Bank, 2004](#)) and the UN Statistics Database ([UNSTAT, 2005](#)). The peatland area fraction at 0.5° resolution is derived from three vector datasets: peatland data in Indonesia and Malaysian Borneo ([Olson et al. 2001](#)); peatland data in Canada ([Tarnocai et al. 2011](#)); and bog, fen and mire data in boreal regions (north of 45°N) outside Canada provided by the Global Lakes and Wetlands Database (GLWD) ([Lehner and Döll, 2004](#)). The climatological peak month for agricultural waste burning is from [van der Werf et al. \(2010\)](#).

Adjustable Parameters and Physical Constants

Values of certain adjustable parameters inherent in the biogeophysical or biogeochemical parameterizations have either been obtained from the literature or calibrated based on comparisons with observations. These are described in the text. Physical constants, generally shared by all of the components in the coupled modeling system, are presented in Table 2.7.

Table 2.7: Physical constants

description	name	value	units
Pi	π	3.14159265358979323846	
Acceleration of gravity	g	9.80616	m s^{-2}
Standard pressure	P_{std}	101325	Pa
Stefan-Boltzmann constant	σ	5.67×10^{-8}	$\text{W m}^{-2} \text{K}^{-4}$
Boltzmann constant	κ	1.38065×10^{-23}	$\text{J K}^{-1} \text{molecule}^{-1}$
Avogadro's number	N_A	6.02214×10^{26}	$\text{molecule kmol}^{-1}$
Universal gas constant	R_{gas}	$N_A \kappa$	$\text{J K}^{-1} \text{kmol}^{-1}$
Molecular weight of dry air	MW_{da}	28.966	kg kmol^{-1}
Dry air gas constant	R_{da}	R_{gas}/MW_{da}	$\text{J K}^{-1} \text{kg}^{-1}$
Molecular weight of water vapor	MW_{wv}	18.016	kg kmol^{-1}
Water vapor gas constant	R_{wv}	R_{gas}/MW_{wv}	$\text{J K}^{-1} \text{kg}^{-1}$
Von Karman constant	k	0.4	-
Freezing temperature of fresh water	T_f	273.15	K
Density of liquid water	ρ_{liq}	1000	kg m^{-3}
Density of ice	ρ_{ice}	917	kg m^{-3}
Specific heat capacity of dry air	C_p	1.00464×10^3	$\text{J kg}^{-1} \text{K}^{-1}$
Specific heat capacity of water	C_{liq}	4.188×10^3	$\text{J kg}^{-1} \text{K}^{-1}$
Specific heat capacity of ice	C_{ice}	2.11727×10^3	$\text{J kg}^{-1} \text{K}^{-1}$
Latent heat of vaporization	λ_{vap}	2.501×10^6	J kg^{-1}
Latent heat of fusion	L_f	3.337×10^5	J kg^{-1}
Latent heat of sublimation	λ_{sub}	$\lambda_{vap} + L_f$	J kg^{-1}
¹ “Thermal conductivity of water”	λ_{liq}	0.57	$\text{W m}^{-1} \text{K}^{-1}$
¹ “Thermal conductivity of ice”	λ_{ice}	2.29	$\text{W m}^{-1} \text{K}^{-1}$
¹ “Thermal conductivity of air”	λ_{air}	0.023 $\text{W m}^{-1} \text{K}^{-1}$	
Radius of the earth	R_e	6.37122	$\times 10^6 \text{ m}$

¹Not shared by other components of the coupled modeling system.

2.3 Surface Albedos

2.3.1 Canopy Radiative Transfer

Radiative transfer within vegetative canopies is calculated from the two-stream approximation of [Dickinson \(1983\)](#) and [Sellers \(1985\)](#) as described by [Bonan \(1996\)](#)

$$-\bar{\mu} \frac{dI \uparrow}{d(L+S)} + [1 - (1 - \beta) \omega] I \uparrow - \omega \beta I \downarrow = \omega \bar{\mu} K \beta_0 e^{-K(L+S)} \quad (2.3)$$

$$\bar{\mu} \frac{dI \downarrow}{d(L+S)} + [1 - (1 - \beta) \omega] I \downarrow - \omega \beta I \uparrow = \omega \bar{\mu} K (1 - \beta_0) e^{-K(L+S)} \quad (2.4)$$

where $I \uparrow$ and $I \downarrow$ are the upward and downward diffuse radiative fluxes per unit incident flux, $K = G(\mu)/\mu$ is the optical depth of direct beam per unit leaf and stem area, μ is the cosine of the zenith angle of the incident beam, $G(\mu)$ is the relative projected area of leaf and stem elements in the direction $\cos^{-1} \mu$, $\bar{\mu}$ is the average inverse diffuse optical depth per unit leaf and stem area, ω is a scattering coefficient, β and β_0 are upscatter parameters for diffuse and direct beam radiation, respectively, L is the exposed leaf area index, and S is the exposed stem area index (section 2.2.1). Given the direct beam albedo $\alpha_{g,\Lambda}^\mu$ and diffuse albedo $\alpha_{g,\Lambda}$ of the ground (section 2.3.2), these equations are solved to calculate the fluxes, per unit incident flux, absorbed by the vegetation, reflected by the vegetation, and

transmitted through the vegetation for direct and diffuse radiation and for visible ($< 0.7\mu\text{m}$) and near-infrared ($\geq 0.7\mu\text{m}$) wavebands. The absorbed radiation is partitioned to sunlit and shaded fractions of the canopy. The optical parameters $G(\mu)$, $\bar{\mu}$, ω , β , and β_0 are calculated based on work in [Sellers \(1985\)](#) as follows.

The relative projected area of leaves and stems in the direction $\cos^{-1}\mu$ is

$$G(\mu) = \phi_1 + \phi_2\mu \quad (2.5)$$

where $\phi_1 = 0.5 - 0.633\chi_L - 0.33\chi_L^2$ and $\phi_2 = 0.877(1 - 2\phi_1)$ for $-0.4 \leq \chi_L \leq 0.6$. χ_L is the departure of leaf angles from a random distribution and equals +1 for horizontal leaves, 0 for random leaves, and -1 for vertical leaves.

The average inverse diffuse optical depth per unit leaf and stem area is

$$\bar{\mu} = \int_0^1 \frac{\mu'}{G(\mu')} d\mu' = \frac{1}{\phi_2} \left[1 - \frac{\phi_1}{\phi_2} \ln \left(\frac{\phi_1 + \phi_2}{\phi_1} \right) \right] \quad (2.6)$$

where μ' is the direction of the scattered flux.

The optical parameters ω , β , and β_0 , which vary with wavelength (Λ), are weighted combinations of values for vegetation and snow, using the canopy snow-covered fraction $f_{can, sno}$ (Chapter 2.7). The optical parameters are

$$\omega_\Lambda = \omega_\Lambda^{veg}(1 - f_{can, sno}) + \omega_\Lambda^{sno}f_{can, sno} \quad (2.7)$$

$$\omega_\Lambda\beta_\Lambda = \omega_\Lambda^{veg}\beta_\Lambda^{veg}(1 - f_{can, sno}) + \omega_\Lambda^{sno}\beta_\Lambda^{sno}f_{can, sno} \quad (2.8)$$

$$\omega_\Lambda\beta_{0,\Lambda} = \omega_\Lambda^{veg}\beta_{0,\Lambda}^{veg}(1 - f_{can, sno}) + \omega_\Lambda^{sno}\beta_{0,\Lambda}^{sno}f_{can, sno} \quad (2.9)$$

The snow and vegetation weights are applied to the products $\omega_\Lambda\beta_\Lambda$ and $\omega_\Lambda\beta_{0,\Lambda}$ because these products are used in the two-stream equations. If there is no snow on the canopy, this reduces to

$$\omega_\Lambda = \omega_\Lambda^{veg} \quad (2.10)$$

$$\omega_\Lambda\beta_\Lambda = \omega_\Lambda^{veg}\beta_\Lambda^{veg} \quad (2.11)$$

$$\omega_\Lambda\beta_{0,\Lambda} = \omega_\Lambda^{veg}\beta_{0,\Lambda}^{veg}. \quad (2.12)$$

For vegetation, $\omega_\Lambda^{veg} = \alpha_\Lambda + \tau_\Lambda$. α_Λ is a weighted combination of the leaf and stem reflectances ($\alpha_\Lambda^{leaf}, \alpha_\Lambda^{stem}$)

$$\alpha_\Lambda = \alpha_\Lambda^{leaf}w_{leaf} + \alpha_\Lambda^{stem}w_{stem} \quad (2.13)$$

where $w_{leaf} = L/(L + S)$ and $w_{stem} = S/(L + S)$. τ_Λ is a weighted combination of the leaf and stem transmittances ($\tau_\Lambda^{leaf}, \tau_\Lambda^{stem}$)

$$\tau_\Lambda = \tau_\Lambda^{leaf}w_{leaf} + \tau_\Lambda^{stem}w_{stem}. \quad (2.14)$$

The upscatter for diffuse radiation is

$$\omega_\Lambda^{veg}\beta_\Lambda^{veg} = \frac{1}{2} [\alpha_\Lambda + \tau_\Lambda + (\alpha_\Lambda - \tau_\Lambda) \cos^2\theta] \quad (2.15)$$

where $\bar{\theta}$ is the mean leaf inclination angle relative to the horizontal plane (i.e., the angle between leaf normal and local vertical) (*Sellers (1985)*). Here, $\cos \bar{\theta}$ is approximated by

$$\cos \bar{\theta} = \frac{1 + \chi_L}{2} \quad (2.16)$$

Using this approximation, for vertical leaves ($\chi_L = -1$, $\bar{\theta} = 90^\circ$), $\omega_\Lambda^{veg} \beta_\Lambda^{veg} = 0.5(\alpha_\Lambda + \tau_\Lambda)$, and for horizontal leaves ($\chi_L = 1$, $\bar{\theta} = 0^\circ$), $\omega_\Lambda^{veg} \beta_\Lambda^{veg} = \alpha_\Lambda$, which agree with both *Dickinson (1983)* and *Sellers (1985)*. For random (spherically distributed) leaves ($\chi_L = 0$, $\bar{\theta} = 60^\circ$), the approximation yields $\omega_\Lambda^{veg} \beta_\Lambda^{veg} = 5/8\alpha_\Lambda + 3/8\tau_\Lambda$ whereas the approximate solution of *Dickinson (1983)* is $\omega_\Lambda^{veg} \beta_\Lambda^{veg} = 2/3\alpha_\Lambda + 1/3\tau_\Lambda$. This discrepancy arises from the fact that a spherical leaf angle distribution has a true mean leaf inclination $\bar{\theta} \approx 57$ (*Campbell and Norman 1998*) in equation , while $\bar{\theta} = 60$ in equation . The upscatter for direct beam radiation is

$$\omega_\Lambda^{veg} \beta_{0,\Lambda}^{veg} = \frac{1 + \bar{\mu}K}{\bar{\mu}K} a_s(\mu)_\Lambda \quad (2.17)$$

where the single scattering albedo is

$$\begin{aligned} a_s(\mu)_\Lambda &= \frac{\omega_\Lambda^{veg}}{2} \int_0^1 \frac{\mu' G(\mu)}{\mu G(\mu') + \mu' G(\mu)} d\mu' \\ &= \frac{\omega_\Lambda^{veg}}{2} \frac{G(\mu)}{\min(\mu\phi_2 + G(\mu), 1e-6)} \left[1 - \frac{\mu\phi_1}{\min(\mu\phi_2 + G(\mu), 1e-6)} \ln \left(\frac{\mu\phi_1 + \min(\mu\phi_2 + G(\mu), 1e-6)}{\mu\phi_1} \right) \right]. \end{aligned} \quad (2.18)$$

Note here the restriction on $\mu\phi_2 + G(\mu)$. We have seen cases where small values can cause unrealistic single scattering albedo associated with the log calculation, thereby eventually causing a negative soil albedo.

The upward diffuse fluxes per unit incident direct beam and diffuse flux (i.e., the surface albedos) are

$$I \uparrow_\Lambda^\mu = \frac{h_1}{\sigma} + h_2 + h_3 \quad (2.19)$$

$$I \uparrow_\Lambda = h_7 + h_8. \quad (2.20)$$

The downward diffuse fluxes per unit incident direct beam and diffuse radiation, respectively, are

$$I \downarrow_\Lambda^\mu = \frac{h_4}{\sigma} e^{-K(L+S)} + h_5 s_1 + \frac{h_6}{s_1} \quad (2.21)$$

$$I \downarrow_\Lambda = h_9 s_1 + \frac{h_{10}}{s_1}. \quad (2.22)$$

With reference to [Figure 2.3](#), the direct beam flux transmitted through the canopy, per unit incident flux, is $e^{-K(L+S)}$, and the direct beam and diffuse fluxes absorbed by the vegetation, per unit incident flux, are

$$\vec{I}_\Lambda^\mu = 1 - I \uparrow_\Lambda^\mu - (1 - \alpha_{g,\Lambda}) I \downarrow_\Lambda^\mu - (1 - \alpha_{g,\Lambda}^\mu) e^{-K(L+S)} \quad (2.23)$$

$$\vec{I}_\Lambda = 1 - I \uparrow_\Lambda - (1 - \alpha_{g,\Lambda}) I \downarrow_\Lambda. \quad (2.24)$$

These fluxes are partitioned to the sunlit and shaded canopy using an analytical solution to the two-stream approximation for sunlit and shaded leaves (*Dai et al. 2004*), as described by *Bonan et al. (2011)*. The absorption of direct beam radiation by sunlit leaves is

$$\vec{I}_{sun,\Lambda}^\mu = (1 - \omega_\Lambda) \left[1 - s_2 + \frac{1}{\bar{\mu}} (a_1 + a_2) \right] \quad (2.25)$$

and for shaded leaves is

$$\vec{I}_{sha,\Lambda}^{\mu} = \vec{I}_{\Lambda}^{\mu} - \vec{I}_{sun,\Lambda}^{\mu} \quad (2.26)$$

with

$$a_1 = \frac{h_1}{\sigma} \left[\frac{1 - s_2^2}{2K} \right] + h_2 \left[\frac{1 - s_2 s_1}{K + h} \right] + h_3 \left[\frac{1 - s_2/s_1}{K - h} \right] \quad (2.27)$$

$$a_2 = \frac{h_4}{\sigma} \left[\frac{1 - s_2^2}{2K} \right] + h_5 \left[\frac{1 - s_2 s_1}{K + h} \right] + h_6 \left[\frac{1 - s_2/s_1}{K - h} \right]. \quad (2.28)$$

For diffuse radiation, the absorbed radiation for sunlit leaves is

$$\vec{I}_{sun,\Lambda} = \left[\frac{1 - \omega_{\Lambda}}{\bar{\mu}} \right] (a_1 + a_2) \quad (2.29)$$

and for shaded leaves is

$$\vec{I}_{sha,\Lambda} = \vec{I}_{\Lambda} - \vec{I}_{sun,\Lambda} \quad (2.30)$$

with

$$a_1 = h_7 \left[\frac{1 - s_2 s_1}{K + h} \right] + h_8 \left[\frac{1 - s_2/s_1}{K - h} \right] \quad (2.31)$$

$$a_2 = h_9 \left[\frac{1 - s_2 s_1}{K + h} \right] + h_{10} \left[\frac{1 - s_2/s_1}{K - h} \right]. \quad (2.32)$$

The parameters $h_1 - h_{10}$, σ , h , s_1 , and s_2 are from *Sellers (1985)* [note the error in h_4 in *Sellers (1985)*]:

$$b = 1 - \omega_{\Lambda} + \omega_{\Lambda} \beta_{\Lambda} \quad (2.33)$$

$$c = \omega_{\Lambda} \beta_{\Lambda} \quad (2.34)$$

$$d = \omega_{\Lambda} \bar{\mu} K \beta_{0,\Lambda} \quad (2.35)$$

$$f = \omega_{\Lambda} \bar{\mu} K (1 - \beta_{0,\Lambda}) \quad (2.36)$$

$$h = \frac{\sqrt{b^2 - c^2}}{\bar{\mu}} \quad (2.37)$$

$$\sigma = (\bar{\mu} K)^2 + c^2 - b^2 \quad (2.38)$$

$$u_1 = b - c / \alpha_{g,\Lambda}^{\mu} \text{ or } u_1 = b - c / \alpha_{g,\Lambda} \quad (2.39)$$

$$u_2 = b - c\alpha_{g,\Lambda}^\mu \text{ or } u_2 = b - c\alpha_{g,\Lambda} \quad (2.40)$$

$$u_3 = f + c\alpha_{g,\Lambda}^\mu \text{ or } u_3 = f + c\alpha_{g,\Lambda} \quad (2.41)$$

$$s_1 = \exp \{-\min [h(L+S), 40]\} \quad (2.42)$$

$$s_2 = \exp \{-\min [K(L+S), 40]\} \quad (2.43)$$

$$p_1 = b + \bar{\mu}h \quad (2.44)$$

$$p_2 = b - \bar{\mu}h \quad (2.45)$$

$$p_3 = b + \bar{\mu}K \quad (2.46)$$

$$p_4 = b - \bar{\mu}K \quad (2.47)$$

$$d_1 = \frac{p_1(u_1 - \bar{\mu}h)}{s_1} - p_2(u_1 + \bar{\mu}h)s_1 \quad (2.48)$$

$$d_2 = \frac{u_2 + \bar{\mu}h}{s_1} - (u_2 - \bar{\mu}h)s_1 \quad (2.49)$$

$$h_1 = -dp_4 - cf \quad (2.50)$$

$$h_2 = \frac{1}{d_1} \left[\left(d - \frac{h_1}{\sigma} p_3 \right) \frac{(u_1 - \bar{\mu}h)}{s_1} - p_2 \left(d - c - \frac{h_1}{\sigma} (u_1 + \bar{\mu}K) \right) s_2 \right] \quad (2.51)$$

$$h_3 = \frac{-1}{d_1} \left[\left(d - \frac{h_1}{\sigma} p_3 \right) (u_1 + \bar{\mu}h) s_1 - p_1 \left(d - c - \frac{h_1}{\sigma} (u_1 + \bar{\mu}K) \right) s_2 \right] \quad (2.52)$$

$$h_4 = -fp_3 - cd \quad (2.53)$$

$$h_5 = \frac{-1}{d_2} \left[\left(\frac{h_4 (u_2 + \bar{\mu}h)}{\sigma s_1} \right) + \left(u_3 - \frac{h_4}{\sigma} (u_2 - \bar{\mu}K) \right) s_2 \right] \quad (2.54)$$

$$h_6 = \frac{1}{d_2} \left[\frac{h_4}{\sigma} (u_2 - \bar{\mu}h) s_1 + \left(u_3 - \frac{h_4}{\sigma} (u_2 - \bar{\mu}K) \right) s_2 \right] \quad (2.55)$$

$$h_7 = \frac{c (u_1 - \bar{\mu}h)}{d_1 s_1} \quad (2.56)$$

$$h_8 = \frac{-c (u_1 + \bar{\mu}h) s_1}{d_1} \quad (2.57)$$

$$h_9 = \frac{u_2 + \bar{\mu}h}{d_2 s_1} \quad (2.58)$$

$$h_{10} = \frac{-s_1 (u_2 - \bar{\mu}h)}{d_2}. \quad (2.59)$$

Plant functional type optical properties (Table 2.8) for trees and shrubs are from [Dorman and Sellers \(1989\)](#). Leaf and stem optical properties (VIS and NIR reflectance and transmittance) were derived for grasslands and crops from full optical range spectra of measured optical properties ([Asner et al. 1998](#)). Optical properties for intercepted snow (Table 2.9) are from [Sellers et al. \(1986\)](#).

Table 2.8: Plant functional type optical properties

Plant Functional Type	χ_L	α_{vis}^{leaf}	α_{nir}^{leaf}	α_{vis}^{stem}	α_{nir}^{stem}	τ_{vis}^{leaf}	τ_{nir}^{leaf}	τ_{vis}^{stem}	τ_{nir}^{stem}
NET Temperate	0.01	0.07	0.35	0.16	0.39	0.05	0.10	0.001	0.001
NET Boreal	0.01	0.07	0.35	0.16	0.39	0.05	0.10	0.001	0.001
NDT Boreal	0.01	0.07	0.35	0.16	0.39	0.05	0.10	0.001	0.001
BET Tropical	0.10	0.10	0.45	0.16	0.39	0.05	0.25	0.001	0.001
BET temperate	0.10	0.10	0.45	0.16	0.39	0.05	0.25	0.001	0.001
BDT tropical	0.01	0.10	0.45	0.16	0.39	0.05	0.25	0.001	0.001
BDT temperate	0.25	0.10	0.45	0.16	0.39	0.05	0.25	0.001	0.001
BDT boreal	0.25	0.10	0.45	0.16	0.39	0.05	0.25	0.001	0.001
BES temperate	0.01	0.07	0.35	0.16	0.39	0.05	0.10	0.001	0.001
BDS temperate	0.25	0.10	0.45	0.16	0.39	0.05	0.25	0.001	0.001
BDS boreal	0.25	0.10	0.45	0.16	0.39	0.05	0.25	0.001	0.001
C ₃ arctic grass	-0.30	0.11	0.35	0.31	0.53	0.05	0.34	0.120	0.250
C ₃ grass	-0.30	0.11	0.35	0.31	0.53	0.05	0.34	0.120	0.250
C ₄ grass	-0.30	0.11	0.35	0.31	0.53	0.05	0.34	0.120	0.250
C ₃ Crop	-0.30	0.11	0.35	0.31	0.53	0.05	0.34	0.120	0.250
Temp Corn	-0.50	0.11	0.35	0.31	0.53	0.05	0.34	0.120	0.250
Spring Wheat	-0.50	0.11	0.35	0.31	0.53	0.05	0.34	0.120	0.250
Temp Soybean	-0.50	0.11	0.35	0.31	0.53	0.05	0.34	0.120	0.250
Cotton	-0.50	0.11	0.35	0.31	0.53	0.05	0.34	0.120	0.250
Rice	-0.50	0.11	0.35	0.31	0.53	0.05	0.34	0.120	0.250
Sugarcane	-0.50	0.11	0.35	0.31	0.53	0.05	0.34	0.120	0.250
Tropical Corn	-0.50	0.11	0.35	0.31	0.53	0.05	0.34	0.120	0.250
Tropical Soybean	-0.50	0.11	0.35	0.31	0.53	0.05	0.34	0.120	0.250

Table 2.9: Intercepted snow optical properties

Parameter	vis	nir
ω^{sno}	0.8	0.4
β^{sno}	0.5	0.5
β_0^{sno}	0.5	0.5

2.3.2 Ground Albedos

The overall direct beam $\alpha_{g, \Lambda}^\mu$ and diffuse $\alpha_{g, \Lambda}$ ground albedos are weighted combinations of “soil” and snow albedos

$$\alpha_{g, \Lambda}^\mu = \alpha_{soi, \Lambda}^\mu (1 - f_{sno}) + \alpha_{sno, \Lambda}^\mu f_{sno} \quad (2.60)$$

$$\alpha_{g, \Lambda} = \alpha_{soi, \Lambda} (1 - f_{sno}) + \alpha_{sno, \Lambda} f_{sno} \quad (2.61)$$

where f_{sno} is the fraction of the ground covered with snow (section 2.8.1).

$\alpha_{soi, \Lambda}^\mu$ and $\alpha_{soi, \Lambda}$ vary with glacier, lake, wetland, and soil surfaces. Glacier albedos are from [Paterson \(1994\)](#)

$$\alpha_{soi, vis}^\mu = \alpha_{soi, vis} = 0.6$$

$$\alpha_{soi, nir}^\mu = \alpha_{soi, nir} = 0.4.$$

Unfrozen lake and wetland albedos depend on the cosine of the solar zenith angle μ

$$\alpha_{soi, \Lambda}^\mu = \alpha_{soi, \Lambda} = 0.05 (\mu + 0.15)^{-1}. \quad (2.62)$$

Frozen lake and wetland albedos are from NCAR LSM ([Bonan 1996](#))

$$\alpha_{soi, vis}^\mu = \alpha_{soi, vis} = 0.60$$

$$\alpha_{soi, nir}^\mu = \alpha_{soi, nir} = 0.40.$$

As in NCAR LSM ([Bonan 1996](#)), soil albedos vary with color class

$$\alpha_{soi, \Lambda}^\mu = \alpha_{soi, \Lambda} = (\alpha_{sat, \Lambda} + \Delta) \leq \alpha_{dry, \Lambda} \quad (2.63)$$

where Δ depends on the volumetric water content of the first soil layer θ_1 (section 2.7.3) as $\Delta = 0.11 - 0.40\theta_1 > 0$, and $\alpha_{sat, \Lambda}$ and $\alpha_{dry, \Lambda}$ are albedos for saturated and dry soil color classes (Table 2.10).

CLM soil colors are prescribed so that they best reproduce observed MODIS local solar noon surface albedo values at the CLM grid cell following the methods of [Lawrence and Chase \(2007\)](#). The soil colors are fitted over the range of 20 soil classes shown in Table 2.10 and compared to the MODIS monthly local solar noon all-sky surface albedo as described in [Strahler et al. \(1999\)](#) and [Schaaf et al. \(2002\)](#). The CLM two-stream radiation model was used to calculate the model equivalent surface albedo using climatological monthly soil moisture along with the vegetation parameters of PFT fraction, LAI, and SAI. The soil color that produced the closest all-sky albedo in the two-stream radiation model was selected as the best fit for the month. The fitted monthly soil colors were averaged over all snow-free months to specify a representative soil color for the grid cell. In cases where there was no snow-free surface albedo for the year, the soil color derived from snow-affected albedo was used to give a representative soil color that included the effects of the minimum permanent snow cover.

Table 2.10: Dry and saturated soil albedos

Color Class	Dry		Saturated		Color Class	Dry		Saturated	
	vis	nir	vis	nir		vis	nir	vis	nir
1	0.36	0.61	0.25	0.50	11	0.24	0.37	0.13	0.26
2	0.34	0.57	0.23	0.46	12	0.23	0.35	0.12	0.24
3	0.32	0.53	0.21	0.42	13	0.22	0.33	0.11	0.22
4	0.31	0.51	0.20	0.40	14	0.20	0.31	0.10	0.20
5	0.30	0.49	0.19	0.38	15	0.18	0.29	0.09	0.18
6	0.29	0.48	0.18	0.36	16	0.16	0.27	0.08	0.16
7	0.28	0.45	0.17	0.34	17	0.14	0.25	0.07	0.14
8	0.27	0.43	0.16	0.32	18	0.12	0.23	0.06	0.12
9	0.26	0.41	0.15	0.30	19	0.10	0.21	0.05	0.10
10	0.25	0.39	0.14	0.28	20	0.08	0.16	0.04	0.08

Snow Albedo

Snow albedo and solar absorption within each snow layer are simulated with the Snow, Ice, and Aerosol Radiative Model (SNICAR), which incorporates a two-stream radiative transfer solution from [Toon et al. \(1989\)](#). Albedo and the vertical absorption profile depend on solar zenith angle, albedo of the substrate underlying snow, mass concentrations of atmospheric-deposited aerosols (black carbon, mineral dust, and organic carbon), and ice effective grain size (r_e), which is simulated with a snow aging routine described in section 2.3.2. Representation of impurity mass concentrations within the snowpack is described in section 2.8.4. Implementation of SNICAR in CLM is also described somewhat by [Flanner and Zender \(2005\)](#) and [Flanner et al. \(2007\)](#).

The two-stream solution requires the following bulk optical properties for each snow layer and spectral band: extinction optical depth (τ), single-scatter albedo (ω), and scattering asymmetry parameter (g). The snow layers used for radiative calculations are identical to snow layers applied elsewhere in CLM, except for the case when snow mass is greater than zero but no snow layers exist. When this occurs, a single radiative layer is specified to have the column snow mass and an effective grain size of freshly-fallen snow (section 2.3.2). The bulk optical properties are weighted functions of each constituent k , computed for each snow layer and spectral band as

$$\tau = \sum_1^k \tau_k \quad (2.64)$$

$$\omega = \frac{\sum_1^k \omega_k \tau_k}{\sum_1^k \tau_k} \quad (2.65)$$

$$g = \frac{\sum_1^k g_k \omega_k \tau_k}{\sum_1^k \omega_k \tau_k} \quad (2.66)$$

For each constituent (ice, two black carbon species, two organic carbon species, and four dust species), ω , g , and the mass extinction cross-section ψ ($\text{m}^2 \text{ kg}^{-1}$) are computed offline with Mie Theory, e.g., applying the computational technique from [Bohren and Huffman \(1983\)](#). The extinction optical depth for each constituent depends on its mass extinction cross-section and layer mass, w_k (kg m^{-1}) as

$$\tau_k = \psi_k w_k \quad (2.67)$$

The two-stream solution ([Toon et al. \(1989\)](#)) applies a tri-diagonal matrix solution to produce upward and downward radiative fluxes at each layer interface, from which net radiation, layer absorption, and surface albedo are easily derived. Solar fluxes are computed in five spectral bands, listed in Table 2.11. Because snow albedo varies strongly

across the solar spectrum, it was determined that four bands were needed to accurately represent the near-infrared (NIR) characteristics of snow, whereas only one band was needed for the visible spectrum. Boundaries of the NIR bands were selected to capture broad radiative features and maximize accuracy and computational efficiency. We partition NIR ($0.7\text{-}5.0 \mu\text{m}$) surface downwelling flux from CLM according to the weights listed in Table 2.11, which are unique for diffuse and direct incident flux. These fixed weights were determined with offline hyperspectral radiative transfer calculations for an atmosphere typical of mid-latitude winter ([Flanner et al. \(2007\)](#)). The tri-diagonal solution includes intermediate terms that allow for easy interchange of two-stream techniques. We apply the Eddington solution for the visible band (following [Wiscombe and Warren 1980](#)) and the hemispheric mean solution ([Toon et al. \(1989\)](#)) for NIR bands. These choices were made because the Eddington scheme works well for highly scattering media, but can produce negative albedo for absorptive NIR bands with diffuse incident flux. Delta scalings are applied to τ , ω , and g ([Wiscombe and Warren 1980](#)) in all spectral bands, producing effective values (denoted with *) that are applied in the two-stream solution

$$\tau^* = (1 - \omega g^2) \tau \quad (2.68)$$

$$\omega^* = \frac{(1 - g^2) \omega}{1 - g^2 \omega} \quad (2.69)$$

$$g^* = \frac{g}{1 + g} \quad (2.70)$$

Table 2.11: Spectral bands and weights used for snow radiative transfer

Spectral band	Direct-beam weight	Diffuse weight
Band 1: $0.3\text{-}0.7 \mu\text{m}$ (visible)	(1.0)	(1.0)
Band 2: $0.7\text{-}1.0 \mu\text{m}$ (near-IR)	0.494	0.586
Band 3: $1.0\text{-}1.2 \mu\text{m}$ (near-IR)	0.181	0.202
Band 4: $1.2\text{-}1.5 \mu\text{m}$ (near-IR)	0.121	0.109
Band 5: $1.5\text{-}5.0 \mu\text{m}$ (near-IR)	0.204	0.103

Under direct-beam conditions, singularities in the radiative approximation are occasionally approached in spectral bands 4 and 5 that produce unrealistic conditions (negative energy absorption in a layer, negative albedo, or total absorbed flux greater than incident flux). When any of these three conditions occur, the Eddington approximation is attempted instead, and if both approximations fail, the cosine of the solar zenith angle is adjusted by 0.02 (conserving incident flux) and a warning message is produced. This situation occurs in only about 1 in 10^6 computations of snow albedo. After looping over the five spectral bands, absorption fluxes and albedo are averaged back into the bulk NIR band used by the rest of CLM.

Soil albedo (or underlying substrate albedo), which is defined for visible and NIR bands, is a required boundary condition for the snow radiative transfer calculation. Currently, the bulk NIR soil albedo is applied to all four NIR snow bands. With ground albedo as a lower boundary condition, SNICAR simulates solar absorption in all snow layers as well as the underlying soil or ground. With a thin snowpack, penetrating solar radiation to the underlying soil can be quite large and heat cannot be released from the soil to the atmosphere in this situation. Thus, if the snowpack has total snow depth less than 0.1 m ($z_{sno} < 0.1$) and there are no explicit snow layers, the solar radiation is absorbed by the top soil layer. If there is a single snow layer, the solar radiation is absorbed in that layer. If there is more than a single snow layer, 75% of the solar radiation is absorbed in the top snow layer, and 25% is absorbed in the next lowest snow layer. This prevents unrealistic soil warming within a single timestep.

The radiative transfer calculation is performed twice for each column containing a mass of snow greater than $1 \times 10^{-30} \text{ kg m}^{-2}$ (excluding lake and urban columns); once each for direct-beam and diffuse incident flux. Absorption in each layer i of pure snow is initially recorded as absorbed flux per unit incident flux on the ground ($S_{sno,i}$), as albedos must be calculated for the next timestep with unknown incident flux. The snow absorption fluxes that are used for

column temperature calculations are

$$S_{g,i} = S_{sno,i} (1 - \alpha_{sno}) \quad (2.71)$$

This weighting is performed for direct-beam and diffuse, visible and NIR fluxes. After the ground-incident fluxes (transmitted through the vegetation canopy) have been calculated for the current time step (sections 2.3.1 and 2.4.1), the layer absorption factors

($S_{g,i}$) are multiplied by the ground-incident fluxes to produce solar absorption (W m^{-2}) in each snow layer and the underlying ground.

Snowpack Optical Properties

Ice optical properties for the five spectral bands are derived offline and stored in a namelist-defined lookup table for online retrieval (see CLM5.0 User's Guide). Mie properties are first computed at fine spectral resolution (470 bands), and are then weighted into the five bands applied by CLM according to incident solar flux, $I^\downarrow(\lambda)$. For example, the broadband mass-extinction cross section ($\bar{\psi}$) over wavelength interval λ_1 to λ_2 is

$$\bar{\psi} = \frac{\int_{\lambda_1}^{\lambda_2} \psi(\lambda) I^\downarrow(\lambda) d\lambda}{\int_{\lambda_1}^{\lambda_2} I^\downarrow(\lambda) d\lambda} \quad (2.72)$$

Broadband single-scatter albedo ($\bar{\omega}$) is additionally weighted by the diffuse albedo for a semi-infinite snowpack (α_{sno})

$$\bar{\omega} = \frac{\int_{\lambda_1}^{\lambda_2} \omega(\lambda) I^\downarrow(\lambda) \alpha_{sno}(\lambda) d\lambda}{\int_{\lambda_1}^{\lambda_2} I^\downarrow(\lambda) \alpha_{sno}(\lambda) d\lambda} \quad (2.73)$$

Inclusion of this additional albedo weight was found to improve accuracy of the five-band albedo solutions (relative to 470-band solutions) because of the strong dependence of optically-thick snowpack albedo on ice grain single-scatter albedo ([Flanner et al. \(2007\)](#)). The lookup tables contain optical properties for lognormal distributions of ice particles over the range of effective radii: $30\mu\text{m} < r_e < 1500\mu\text{m}$, at $1\mu\text{m}$ resolution. Single-scatter albedos for the end-members of this size range are listed in [Table 2.12](#).

Optical properties for black carbon are described in [Flanner et al. \(2007\)](#). Single-scatter albedo, mass extinction cross-section, and asymmetry parameter values for all snowpack species, in the five spectral bands used, are listed in [Table 2.12](#), [Table 2.13](#), and [Table 2.14](#). These properties were also derived with Mie Theory, using various published sources of indices of refraction and assumptions about particle size distribution. Weighting into the five CLM spectral bands was determined only with incident solar flux, as in equation .

Table 2.12: Single-scatter albedo values used for snowpack impurities and ice

Species	Band 1	Band 2	Band 3	Band 4	Band 5
Hydrophilic black carbon	0.516	0.434	0.346	0.276	0.139
Hydrophobic black carbon	0.288	0.187	0.123	0.089	0.040
Hydrophilic organic carbon	0.997	0.994	0.990	0.987	0.951
Hydrophobic organic carbon	0.963	0.921	0.860	0.814	0.744
Dust 1	0.979	0.994	0.993	0.993	0.953
Dust 2	0.944	0.984	0.989	0.992	0.983
Dust 3	0.904	0.965	0.969	0.973	0.978
Dust 4	0.850	0.940	0.948	0.953	0.955
Ice ($r_e = 30\mu\text{m}$)	0.9999	0.9999	0.9992	0.9938	0.9413
Ice ($r_e = 1500\mu\text{m}$)	0.9998	0.9960	0.9680	0.8730	0.5500

Table 2.13: Mass extinction values ($\text{m}^2 \text{ kg}^{-1}$) used for snowpack impurities and ice

Species	Band 1	Band 2	Band 3	Band 4	Band 5
Hydrophilic black carbon	25369	12520	7739	5744	3527
Hydrophobic black carbon	11398	5923	4040	3262	2224
Hydrophilic organic carbon	37774	22112	14719	10940	5441
Hydrophobic organic carbon	3289	1486	872	606	248
Dust 1	2687	2420	1628	1138	466
Dust 2	841	987	1184	1267	993
Dust 3	388	419	400	397	503
Dust 4	197	203	208	205	229
Ice ($r_e = 30 \mu\text{m}$)	55.7	56.1	56.3	56.6	57.3
Ice ($r_e = 1500 \mu\text{m}$)	1.09	1.09	1.09	1.09	1.1

Table 2.14: Asymmetry scattering parameters used for snowpack impurities and ice.

Species	Band 1	Band 2	Band 3	Band 4	Band 5
Hydrophilic black carbon	0.52	0.34	0.24	0.19	0.10
Hydrophobic black carbon	0.35	0.21	0.15	0.11	0.06
Hydrophilic organic carbon	0.77	0.75	0.72	0.70	0.64
Hydrophobic organic carbon	0.62	0.57	0.54	0.51	0.44
Dust 1	0.69	0.72	0.67	0.61	0.44
Dust 2	0.70	0.65	0.70	0.72	0.70
Dust 3	0.79	0.75	0.68	0.63	0.67
Dust 4	0.83	0.79	0.77	0.76	0.73
Ice ($r_e = 30 \mu\text{m}$)	0.88	0.88	0.88	0.88	0.90
Ice ($r_e = 1500 \mu\text{m}$)	0.89	0.90	0.90	0.92	0.97

Snow Aging

Snow aging is represented as evolution of the ice effective grain size (r_e). Previous studies have shown that use of spheres which conserve the surface area-to-volume ratio (or specific surface area) of ice media composed of more complex shapes produces relatively small errors in simulated hemispheric fluxes (e.g., [Grenfell and Warren 1999](#)). Effective radius is the surface area-weighted mean radius of an ensemble of spherical particles and is directly related to specific surface area (SSA) as $r_e = 3 / (\rho_{ice} SSA)$, where ρ_{ice} is the density of ice. Hence, r_e is a simple and practical metric for relating the snowpack microphysical state to dry snow radiative characteristics.

Wet snow processes can also drive rapid changes in albedo. The presence of liquid water induces rapid coarsening of the surrounding ice grains (e.g., [Brun 1989](#)), and liquid water tends to refreeze into large ice clumps that darken the bulk snowpack. The presence of small liquid drops, by itself, does not significantly darken snowpack, as ice and water have very similar indices of refraction throughout the solar spectrum. Pooled or ponded water, however, can significantly darken snowpack by greatly reducing the number of refraction events per unit mass. This influence is not currently accounted for.

The net change in effective grain size occurring each time step is represented in each snow layer as a summation of changes caused by dry snow metamorphism ($dr_{e,dry}$), liquid water-induced metamorphism ($dr_{e,wet}$), refreezing of liquid water, and addition of freshly-fallen snow. The mass of each snow layer is partitioned into fractions of snow carrying over from the previous time step (f_{old}), freshly-fallen snow (f_{new}), and refrozen liquid water (f_{rfz}), such that snow r_e is updated each time step t as

$$r_e(t) = [r_e(t-1) + dr_{e,dry} + dr_{e,wet}] f_{old} + r_{e,0} f_{new} + r_{e,rfz} f_{rfz} \quad (2.74)$$

Here, the effective radius of freshly-fallen snow ($r_{e,0}$) is fixed globally at $54.5 \mu\text{m}$ (corresponding to a specific surface area of $60 \text{ m}^2 \text{ kg}^{-1}$), and the effective radius of refrozen liquid water ($r_{e,rfz}$) is set to $1000 \mu\text{m}$.

Dry snow aging is based on a microphysical model described by [Flanner and Zender \(2006\)](#). This model simulates diffusive vapor flux amongst collections of ice crystals with various size and inter-particle spacing. Specific surface area and effective radius are prognosed for any combination of snow temperature, temperature gradient, density, and initial size distribution. The combination of warm snow, large temperature gradient, and low density produces the most rapid snow aging, whereas aging proceeds slowly in cold snow, regardless of temperature gradient and density. Because this model is currently too computationally expensive for inclusion in climate models, we fit parametric curves to model output over a wide range of snow conditions and apply these parameters in CLM. The functional form of the parametric equation is

$$\frac{dr_{e,dry}}{dt} = \left(\frac{dr_e}{dt} \right)_0 \left(\frac{\eta}{(r_e - r_{e,0}) + \eta} \right)^{1/\kappa} \quad (2.75)$$

The parameters $(\frac{dr_e}{dt})_0$, η , and κ are retrieved interactively from a lookup table with dimensions corresponding to snow temperature, temperature gradient, and density. The domain covered by this lookup table includes temperature ranging from 223 to 273 K, temperature gradient ranging from 0 to 300 K m^{-1} , and density ranging from 50 to 400 kg m^{-3} . Temperature gradient is calculated at the midpoint of each snow layer n , using mid-layer temperatures (T_n) and snow layer thicknesses (dz_n), as

$$\left(\frac{dT}{dz} \right)_n = \frac{1}{dz_n} \text{abs} \left[\frac{T_{n-1}dz_n + T_n dz_{n-1}}{dz_n + dz_{n-1}} + \frac{T_{n+1}dz_n + T_n dz_{n+1}}{dz_n + dz_{n+1}} \right] \quad (2.76)$$

For the bottom snow layer ($n = 0$), T_{n+1} is taken as the temperature of the top soil layer, and for the top snow layer it is assumed that $T_{n-1} = T_n$.

The contribution of liquid water to enhanced metamorphism is based on parametric equations published by [Brun \(1989\)](#), who measured grain growth rates under different liquid water contents. This relationship, expressed in terms of $r_e (\mu\text{m})$ and subtracting an offset due to dry aging, depends on the mass liquid water fraction f_{liq} as

$$\frac{dr_e}{dt} = \frac{10^{18} C_1 f_{liq}^3}{4\pi r_e^2} \quad (2.77)$$

The constant C_1 is 4.22×10^{-13} , and: $f_{liq} = w_{liq}/(w_{liq} + w_{ice})$ (Chapter 2.8).

In cases where snow mass is greater than zero, but a snow layer has not yet been defined, r_e is set to $r_{e,0}$. When snow layers are combined or divided, r_e is calculated as a mass-weighted mean of the two layers, following computations of other state variables (section 2.8.7). Finally, the allowable range of r_e , corresponding to the range over which Mie optical properties have been defined, is $30\text{-}1500 \mu\text{m}$.

2.3.3 Solar Zenith Angle

The CLM uses the same formulation for solar zenith angle as the Community Atmosphere Model. The cosine of the solar zenith angle μ is

$$\mu = \sin \phi \sin \delta - \cos \phi \cos \delta \cos h \quad (2.78)$$

where h is the solar hour angle (radians) (24 hour periodicity), δ is the solar declination angle (radians), and ϕ is latitude (radians) (positive in Northern Hemisphere). The solar hour angle h (radians) is

$$h = 2\pi d + \theta \quad (2.79)$$

where d is calendar day ($d = 0.0$ at 0Z on January 1), and θ is longitude (radians) (positive east of the Greenwich meridian).

The solar declination angle δ is calculated as in *Berger (1978a,b)* and is valid for one million years past or hence, relative to 1950 A.D. The orbital parameters may be specified directly or the orbital parameters are calculated for the desired year. The required orbital parameters to be input by the user are the obliquity of the Earth ε (degrees, $-90^\circ < \varepsilon < 90^\circ$), Earth's eccentricity e ($0.0 < e < 0.1$), and the longitude of the perihelion relative to the moving vernal equinox $\tilde{\omega}$ ($0^\circ < \tilde{\omega} < 360^\circ$) (unadjusted for the apparent orbit of the Sun around the Earth (*Berger et al. 1993*)). The solar declination δ (radians) is

$$\delta = \sin^{-1} [\sin(\varepsilon) \sin(\lambda)] \quad (2.80)$$

where ε is Earth's obliquity and λ is the true longitude of the Earth.

The obliquity of the Earth ε (degrees) is

$$\varepsilon = \varepsilon * + \sum_{i=1}^{i=47} A_i \cos(f_i t + \delta_i) \quad (2.81)$$

where $\varepsilon*$ is a constant of integration (Table 2.15), A_i , f_i , and δ_i are amplitude, mean rate, and phase terms in the cosine series expansion (*Berger (1978a,b)*), and $t = t_0 - 1950$ where t_0 is the year. The series expansion terms are not shown here but can be found in the source code file shr_orb_mod.F90.

The true longitude of the Earth λ (radians) is counted counterclockwise from the vernal equinox ($\lambda = 0$ at the vernal equinox)

$$\lambda = \lambda_m + \left(2e - \frac{1}{4}e^3\right) \sin(\lambda_m - \tilde{\omega}) + \frac{5}{4}e^2 \sin 2(\lambda_m - \tilde{\omega}) + \frac{13}{12}e^3 \sin 3(\lambda_m - \tilde{\omega}) \quad (2.82)$$

where λ_m is the mean longitude of the Earth at the vernal equinox, e is Earth's eccentricity, and $\tilde{\omega}$ is the longitude of the perihelion relative to the moving vernal equinox. The mean longitude λ_m is

$$\lambda_m = \lambda_{m0} + \frac{2\pi(d - d_{ve})}{365} \quad (2.83)$$

where $d_{ve} = 80.5$ is the calendar day at vernal equinox (March 21 at noon), and

$$\lambda_{m0} = 2 \left[\left(\frac{1}{2}e + \frac{1}{8}e^3 \right) (1 + \beta) \sin \tilde{\omega} - \frac{1}{4}e^2 \left(\frac{1}{2} + \beta \right) \sin 2\tilde{\omega} + \frac{1}{8}e^3 \left(\frac{1}{3} + \beta \right) \sin 3\tilde{\omega} \right] \quad (2.84)$$

where $\beta = \sqrt{1 - e^2}$. Earth's eccentricity e is

$$e = \sqrt{(e^{\cos})^2 + (e^{\sin})^2} \quad (2.85)$$

where

$$\begin{aligned} e^{\cos} &= \sum_{j=1}^{j=19} M_j \cos(g_j t + B_j), \\ e^{\sin} &= \sum_{j=1}^{j=19} M_j \sin(g_j t + B_j) \end{aligned} \quad (2.86)$$

are the cosine and sine series expansions for e , and M_j , g_j , and B_j are amplitude, mean rate, and phase terms in the series expansions (*Berger (1978a,b)*). The longitude of the perihelion relative to the moving vernal equinox $\tilde{\omega}$ (degrees) is

$$\tilde{\omega} = \Pi \frac{180}{\pi} + \psi \quad (2.87)$$

where Π is the longitude of the perihelion measured from the reference vernal equinox (i.e., the vernal equinox at 1950 A.D.) and describes the absolute motion of the perihelion relative to the fixed stars, and ψ is the annual general precession in longitude and describes the absolute motion of the vernal equinox along Earth's orbit relative to the fixed stars. The general precession ψ (degrees) is

$$\psi = \frac{\tilde{\psi}t}{3600} + \zeta + \sum_{i=1}^{i=78} F_i \sin(f'_i t + \delta'_i) \quad (2.88)$$

where $\tilde{\psi}$ (arcseconds) and ζ (degrees) are constants (Table 2.15), and F_i , f'_i , and δ'_i are amplitude, mean rate, and phase terms in the sine series expansion ([Berger \(1978a,b\)](#)). The longitude of the perihelion Π (radians) depends on the sine and cosine series expansions for the eccentricity e as follows:

$$\Pi = \left\{ \begin{array}{ll} 0 & \text{for } -1 \times 10^{-8} \leq e^{\cos} \leq 1 \times 10^{-8} \text{ and } e^{\sin} = 0 \\ 1.5\pi & \text{for } -1 \times 10^{-8} \leq e^{\cos} \leq 1 \times 10^{-8} \text{ and } e^{\sin} < 0 \\ 0.5\pi & \text{for } -1 \times 10^{-8} \leq e^{\cos} \leq 1 \times 10^{-8} \text{ and } e^{\sin} > 0 \\ \tan^{-1} \left[\frac{e^{\sin}}{e^{\cos}} \right] + \pi & \text{for } e^{\cos} < -1 \times 10^{-8} \\ \tan^{-1} \left[\frac{e^{\sin}}{e^{\cos}} \right] + 2\pi & \text{for } e^{\cos} > 1 \times 10^{-8} \text{ and } e^{\sin} < 0 \\ \tan^{-1} \left[\frac{e^{\sin}}{e^{\cos}} \right] & \text{for } e^{\cos} > 1 \times 10^{-8} \text{ and } e^{\sin} \geq 0 \end{array} \right\}. \quad (2.89)$$

The numerical solution for the longitude of the perihelion $\tilde{\omega}$ is constrained to be between 0 and 360 degrees (measured from the autumn equinox). A constant 180 degrees is then added to $\tilde{\omega}$ because the Sun is considered as revolving around the Earth (geocentric coordinate system) ([Berger et al. 1993](#)).

Table 2.15: Orbital parameters

Parameter	
ε^*	23.320556
$\tilde{\psi}$ (arcseconds)	50.439273
ζ (degrees)	3.392506

2.4 Radiative Fluxes

The net radiation at the surface is $(\vec{S}_v + \vec{S}_g) - (\vec{L}_v + \vec{L}_g)$, where \vec{S} is the net solar flux absorbed by the vegetation (“v”) and the ground (“g”) and \vec{L} is the net longwave flux (positive toward the atmosphere) (W m^{-2}).

2.4.1 Solar Fluxes

[Figure 2.3](#) illustrates the direct beam and diffuse fluxes in the canopy.

$I \uparrow_{\Lambda}^{\mu}$ and $I \uparrow_{\Lambda}$ are the upward diffuse fluxes, per unit incident direct beam and diffuse flux (section 2.3.1). $I \downarrow_{\Lambda}^{\mu}$ and $I \downarrow_{\Lambda}$ are the downward diffuse fluxes below the vegetation per unit incident direct beam and diffuse radiation (section 2.3.1). The direct beam flux transmitted through the canopy, per unit incident flux, is $e^{-K(L+S)}$. \vec{I}_{Λ}^{μ} and \vec{I}_{Λ} are the fluxes absorbed by the vegetation, per unit incident direct beam and diffuse radiation (section 2.3.1). $\alpha_{g, \Lambda}^{\mu}$ and $\alpha_{g, \Lambda}$ are the direct beam and diffuse ground albedos (section 2.3.2). L and S are the exposed leaf area index and stem area index (section 2.2.1). K is the optical depth of direct beam per unit leaf and stem area (section 2.3.1).

For clarity, terms involving $T^{n+1} - T^n$ are not shown in (c).

The total solar radiation absorbed by the vegetation and ground is

$$\vec{S}_v = \sum_{\Lambda} S_{atm} \downarrow_{\Lambda}^{\mu} \vec{I}_{\Lambda}^{\mu} + S_{atm} \downarrow_{\Lambda} \vec{I}_{\Lambda} \quad (2.90)$$

$$\vec{S}_g = \sum_{\Lambda} S_{atm} \downarrow_{\Lambda}^{\mu} e^{-K(L+S)} \left(1 - \alpha_{g, \Lambda}^{\mu} \right) + (S_{atm} \downarrow_{\Lambda}^{\mu} I \downarrow_{\Lambda}^{\mu} + S_{atm} \downarrow_{\Lambda} I \downarrow_{\Lambda}) (1 - \alpha_{g, \Lambda}) \quad (2.91)$$

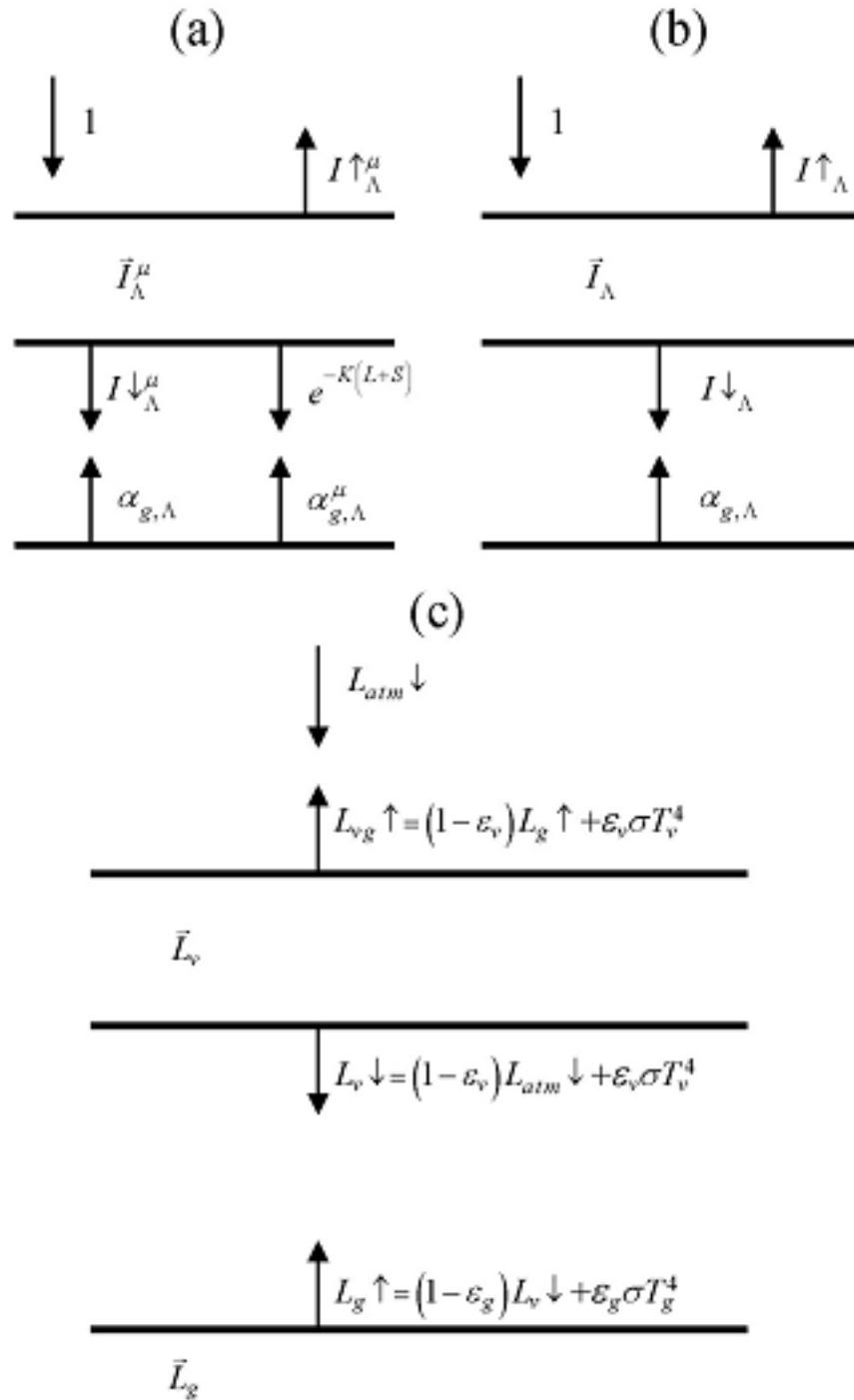


Figure 2.3: Schematic diagram of (a) direct beam radiation, (b) diffuse solar radiation, and (c) longwave radiation absorbed, transmitted, and reflected by vegetation and ground.

where $S_{atm} \downarrow_{\Lambda}^{\mu}$ and $S_{atm} \downarrow_{\Lambda}$ are the incident direct beam and diffuse solar fluxes (W m^{-2}). For non-vegetated surfaces, $e^{-K(L+S)} = 1$, $\vec{I}_{\Lambda}^{\mu} = \vec{I}_{\Lambda} = 0$, $I \downarrow_{\Lambda}^{\mu} = 0$, and $I \downarrow_{\Lambda} = 1$, so that

$$\begin{aligned}\vec{S}_g &= \sum_{\Lambda} S_{atm} \downarrow_{\Lambda}^{\mu} \left(1 - \alpha_{g,\Lambda}^{\mu}\right) + S_{atm} \downarrow_{\Lambda} (1 - \alpha_{g,\Lambda}) \\ \vec{S}_v &= 0\end{aligned}. \quad (2.92)$$

Solar radiation is conserved as

$$\sum_{\Lambda} (S_{atm} \downarrow_{\Lambda}^{\mu} + S_{atm} \downarrow_{\Lambda}) = (\vec{S}_v + \vec{S}_g) + \sum_{\Lambda} (S_{atm} \downarrow_{\Lambda}^{\mu} I \uparrow_{\Lambda}^{\mu} + S_{atm} \downarrow_{\Lambda} I \uparrow_{\Lambda}) \quad (2.93)$$

where the latter term in parentheses is reflected solar radiation.

Photosynthesis and transpiration depend non-linearly on solar radiation, via the light response of stomata. The canopy is treated as two leaves (sunlit and shaded) and the solar radiation in the visible waveband ($< 0.7 \mu\text{m}$) absorbed by the vegetation is apportioned to the sunlit and shaded leaves (section 2.3.1). The absorbed photosynthetically active (visible waveband) radiation averaged over the sunlit canopy (per unit plant area) is

$$\phi^{sun} = \left(\vec{I}_{sun,vis}^{\mu} S_{atm} \downarrow_{vis}^{\mu} + \vec{I}_{sun,vis} S_{atm} \downarrow_{vis} \right) / L^{sun} \quad (2.94)$$

and the absorbed radiation for the average shaded leaf (per unit plant area) is

$$\phi^{sha} = \left(\vec{I}_{sha,vis}^{\mu} S_{atm} \downarrow_{vis}^{\mu} + \vec{I}_{sha,vis} S_{atm} \downarrow_{vis} \right) / L^{sha} \quad (2.95)$$

with L^{sun} and L^{sha} the sunlit and shaded plant area index, respectively. The sunlit plant area index is

$$L^{sun} = \frac{1 - e^{-K(L+S)}}{K} \quad (2.96)$$

and the shaded leaf area index is $L^{sha} = (L + S) - L^{sun}$. In calculating L^{sun} ,

$$K = \frac{G(\mu)}{\mu} \quad (2.97)$$

where $G(\mu)$ and μ are parameters in the two-stream approximation (section 2.3.1).

The model uses the two-stream approximation to calculate radiative transfer of direct and diffuse radiation through a canopy that is differentiated into leaves that are sunlit and those that are shaded (section 2.3.1). The two-stream equations are integrated over all plant area (leaf and stem area) in the canopy. The model has an optional (though not supported) multi-layer canopy, as described by [Bonan et al. \(2012\)](#). The multi-layer model is only intended to address the non-linearity of light profiles, photosynthesis, and stomatal conductance in the plant canopy.

In the multi-layer canopy, canopy-integrated radiative fluxes are calculated from the two-stream approximation. The model additionally derives the light profile with depth in the canopy by taking the derivatives of the absorbed radiative fluxes with respect to plant area index ($L' = L + S$) and evaluating them incrementally through the canopy with cumulative plant area index (x). The terms $d\vec{I}_{sun,\Lambda}^{\mu}(x)/dL'$ and $d\vec{I}_{sun,\Lambda}(x)/dL'$ are the direct beam and diffuse solar radiation, respectively, absorbed by the sunlit fraction of the canopy (per unit plant area) at a depth defined by the cumulative plant area index x ; $d\vec{I}_{sha,\Lambda}^{\mu}(x)/dL'$ and $d\vec{I}_{sha,\Lambda}(x)/dL'$ are the corresponding fluxes for the shaded fraction of the canopy at depth x . These fluxes are normalized by the sunlit or shaded fraction at depth x , defined by $f_{sun} = \exp(-Kx)$, to give fluxes per unit sunlit or shaded plant area at depth x .

2.4.2 Longwave Fluxes

The net longwave radiation (W m^{-2}) (positive toward the atmosphere) at the surface is

$$\vec{L} = L \uparrow - L_{atm} \downarrow \quad (2.98)$$

where $L \uparrow$ is the upward longwave radiation from the surface and $L_{atm} \downarrow$ is the downward atmospheric longwave radiation (W m^{-2}). The radiative temperature T_{rad} (K) is defined from the upward longwave radiation as

$$T_{rad} = \left(\frac{L \uparrow}{\sigma} \right)^{1/4} \quad (2.99)$$

where σ is the Stefan-Boltzmann constant ($\text{W m}^{-2} \text{ K}^{-4}$) (Table 2.7). With reference to Figure 2.3, the upward longwave radiation from the surface to the atmosphere is

$$\begin{aligned} L \uparrow &= \delta_{veg} L_{vg} \uparrow + (1 - \delta_{veg}) (1 - \varepsilon_g) L_{atm} \downarrow + \\ &\quad (1 - \delta_{veg}) \varepsilon_g \sigma (T_g^n)^4 + 4 \varepsilon_g \sigma (T_g^n)^3 (T_g^{n+1} - T_g^n) \end{aligned} \quad (2.100)$$

where $L_{vg} \uparrow$ is the upward longwave radiation from the vegetation/soil system for exposed leaf and stem area $L + S \geq 0.05$, δ_{veg} is a step function and is zero for $L + S < 0.05$ and one otherwise, ε_g is the ground emissivity, and T_g^{n+1} and T_g^n are the snow/soil surface temperatures at the current and previous time steps, respectively (Soil and Snow Temperatures).

For non-vegetated surfaces, the above equation reduces to

$$L \uparrow = (1 - \varepsilon_g) L_{atm} \downarrow + \varepsilon_g \sigma (T_g^n)^4 + 4 \varepsilon_g \sigma (T_g^n)^3 (T_g^{n+1} - T_g^n) \quad (2.101)$$

where the first term is the atmospheric longwave radiation reflected by the ground, the second term is the longwave radiation emitted by the ground, and the last term is the increase (decrease) in longwave radiation emitted by the ground due to an increase (decrease) in ground temperature.

For vegetated surfaces, the upward longwave radiation from the surface reduces to

$$L \uparrow = L_{vg} \uparrow + 4 \varepsilon_g \sigma (T_g^n)^3 (T_g^{n+1} - T_g^n) \quad (2.102)$$

where

$$\begin{aligned} L_{vg} \uparrow &= (1 - \varepsilon_g) (1 - \varepsilon_v) (1 - \varepsilon_v) L_{atm} \downarrow \\ &\quad + \varepsilon_v [1 + (1 - \varepsilon_g) (1 - \varepsilon_v)] \sigma (T_v^n)^3 [T_v^n + 4 (T_v^{n+1} - T_v^n)] \\ &\quad + \varepsilon_g (1 - \varepsilon_v) \sigma (T_g^n)^4 \\ &= (1 - \varepsilon_g) (1 - \varepsilon_v) (1 - \varepsilon_v) L_{atm} \downarrow \\ &\quad + \varepsilon_v \sigma (T_v^n)^4 \\ &\quad + \varepsilon_v (1 - \varepsilon_g) (1 - \varepsilon_v) \sigma (T_v^n)^4 \\ &\quad + 4 \varepsilon_v \sigma (T_v^n)^3 (T_v^{n+1} - T_v^n) \\ &\quad + 4 \varepsilon_v (1 - \varepsilon_g) (1 - \varepsilon_v) \sigma (T_v^n)^3 (T_v^{n+1} - T_v^n) \\ &\quad + \varepsilon_g (1 - \varepsilon_v) \sigma (T_g^n)^4 \end{aligned} \quad (2.103)$$

where ε_v is the vegetation emissivity and T_v^{n+1} and T_v^n are the vegetation temperatures at the current and previous time steps, respectively (Momentum, Sensible Heat, and Latent Heat Fluxes). The first term in the equation above is the atmospheric longwave radiation that is transmitted through the canopy, reflected by the ground, and transmitted through the canopy to the atmosphere. The second term is the longwave radiation emitted by the canopy directly to the atmosphere. The third term is the longwave radiation emitted downward from the canopy, reflected by the ground, and transmitted through the canopy to the atmosphere. The fourth term is the increase (decrease) in longwave radiation due to an increase (decrease) in canopy temperature that is emitted by the canopy directly to the atmosphere. The fifth term is the increase (decrease) in longwave radiation due to an increase (decrease) in canopy temperature that is emitted downward from the canopy, reflected from the ground, and transmitted through the canopy to the atmosphere. The last term is the longwave radiation emitted by the ground and transmitted through the canopy to the atmosphere.

The upward longwave radiation from the ground is

$$L_g \uparrow = (1 - \varepsilon_g) L_v \downarrow + \varepsilon_g \sigma (T_g^n)^4 \quad (2.104)$$

where $L_v \downarrow$ is the downward longwave radiation below the vegetation

$$L_v \downarrow = (1 - \varepsilon_v) L_{atm} \downarrow + \varepsilon_v \sigma (T_v^n)^4 + 4\varepsilon_v \sigma (T_v^n)^3 (T_v^{n+1} - T_v^n). \quad (2.105)$$

The net longwave radiation flux for the ground is (positive toward the atmosphere)

$$\vec{L}_g = \varepsilon_g \sigma (T_g^n)^4 - \delta_{veg} \varepsilon_g L_v \downarrow - (1 - \delta_{veg}) \varepsilon_g L_{atm} \downarrow. \quad (2.106)$$

The above expression for \vec{L}_g is the net longwave radiation forcing that is used in the soil temperature calculation (*Soil and Snow Temperatures*). Once updated soil temperatures have been obtained, the term $4\varepsilon_g \sigma (T_g^n)^3 (T_g^{n+1} - T_g^n)$ is added to \vec{L}_g to calculate the ground heat flux (section 2.5.4)

The net longwave radiation flux for vegetation is (positive toward the atmosphere)

$$\vec{L}_v = [2 - \varepsilon_v (1 - \varepsilon_g)] \varepsilon_v \sigma (T_v)^4 - \varepsilon_v \varepsilon_g \sigma (T_g^n)^4 - \varepsilon_v [1 + (1 - \varepsilon_g) (1 - \varepsilon_v)] L_{atm} \downarrow. \quad (2.107)$$

These equations assume that absorptivity equals emissivity. The emissivity of the ground is

$$\varepsilon_g = \varepsilon_{soi} (1 - f_{sno}) + \varepsilon_{sno} f_{sno} \quad (2.108)$$

where $\varepsilon_{soi} = 0.96$ for soil, 0.97 for glacier, and 0.96 for wetland, $\varepsilon_{sno} = 0.97$, and f_{sno} is the fraction of ground covered by snow (section 2.8.1). The vegetation emissivity is

$$\varepsilon_v = 1 - e^{-(L+S)/\bar{\mu}} \quad (2.109)$$

where L and S are the leaf and stem area indices (section 2.2.1) and $\bar{\mu} = 1$ is the average inverse optical depth for longwave radiation.

2.5 Momentum, Sensible Heat, and Latent Heat Fluxes

The zonal τ_x and meridional τ_y momentum fluxes ($\text{kg m}^{-1} \text{s}^{-2}$), sensible heat flux H (W m^{-2}), and water vapor flux E ($\text{kg m}^{-2} \text{s}^{-1}$) between the atmosphere at reference height $z_{atm,x}$ (m) [where x is height for wind (momentum) (m), temperature (sensible heat) (h), and humidity (water vapor) (w); with zonal and meridional winds u_{atm} and v_{atm} (m s^{-1}), potential temperature θ_{atm} (K), and specific humidity q_{atm} (kg kg^{-1})] and the surface [with u_s , v_s , θ_s , and q_s] are

$$\tau_x = -\rho_{atm} \frac{(u_{atm} - u_s)}{r_{am}} \quad (2.110)$$

$$\tau_y = -\rho_{atm} \frac{(v_{atm} - v_s)}{r_{am}} \quad (2.111)$$

$$H = -\rho_{atm} C_p \frac{(\theta_{atm} - \theta_s)}{r_{ah}} \quad (2.112)$$

$$E = -\rho_{atm} \frac{(q_{atm} - q_s)}{r_{aw}}. \quad (2.113)$$

These fluxes are derived in the next section from Monin-Obukhov similarity theory developed for the surface layer (i.e., the nearly constant flux layer above the surface sublayer). In this derivation, u_s and v_s are defined to equal zero

at height $z_{0m} + d$ (the apparent sink for momentum) so that r_{am} is the aerodynamic resistance (s m^{-1}) for momentum between the atmosphere at height $z_{atm,m}$ and the surface at height $z_{0m} + d$. Thus, the momentum fluxes become

$$\tau_x = -\rho_{atm} \frac{u_{atm}}{r_{am}} \quad (2.114)$$

$$\tau_y = -\rho_{atm} \frac{v_{atm}}{r_{am}}. \quad (2.115)$$

Likewise, θ_s and q_s are defined at heights $z_{0h} + d$ and $z_{0w} + d$ (the apparent sinks for heat and water vapor, respectively) r_{aw} are the aerodynamic resistances (s m^{-1}) to sensible heat and water vapor transfer between the atmosphere at heights $z_{atm,h}$ and $z_{atm,w}$ and the surface at heights $z_{0h} + d$ and $z_{0w} + d$, respectively. The specific heat capacity of air C_p ($\text{J kg}^{-1} \text{ K}^{-1}$) is a constant (Table 2.7). The atmospheric potential temperature used here is

$$\theta_{atm} = T_{atm} + \Gamma_d z_{atm,h} \quad (2.116)$$

where T_{atm} is the air temperature (K) at height $z_{atm,h}$ and $\Gamma_d = 0.0098 \text{ K m}^{-1}$ is the negative of the dry adiabatic lapse rate [this expression is first-order equivalent to $\theta_{atm} = T_{atm} (P_{srf}/P_{atm})^{R_{da}/C_p}$ (Stull 1988), where P_{srf} is the surface pressure (Pa), P_{atm} is the atmospheric pressure (Pa), and R_{da} is the gas constant for dry air ($\text{J kg}^{-1} \text{ K}^{-1}$) (Table 2.7)]. By definition, $\theta_s = T_s$. The density of moist air (kg m^{-3}) is

$$\rho_{atm} = \frac{P_{atm} - 0.378e_{atm}}{R_{da}T_{atm}} \quad (2.117)$$

where the atmospheric vapor pressure e_{atm} (Pa) is derived from the atmospheric specific humidity q_{atm}

$$e_{atm} = \frac{q_{atm} P_{atm}}{0.622 + 0.378q_{atm}}. \quad (2.118)$$

2.5.1 Monin-Obukhov Similarity Theory

The surface vertical kinematic fluxes of momentum $\overline{u'w'}$ and $\overline{v'w'}$ ($\text{m}^2 \text{ s}^{-2}$), sensible heat $\overline{\theta'w'}$ (K m s^{-1}), and latent heat $\overline{q'w'}$ ($\text{kg kg}^{-1} \text{ m s}^{-1}$), where u' , v' , w' , θ' , and q' are zonal horizontal wind, meridional horizontal wind, vertical velocity, potential temperature, and specific humidity turbulent fluctuations about the mean, are defined from Monin-Obukhov similarity applied to the surface layer. This theory states that when scaled appropriately, the dimensionless mean horizontal wind speed, mean potential temperature, and mean specific humidity profile gradients depend on unique functions of $\zeta = \frac{z-d}{L}$ (Zeng et al. 1998) as

$$\frac{k(z-d)}{u_*} \frac{\partial |u|}{\partial z} = \phi_m(\zeta) \quad (2.119)$$

$$\frac{k(z-d)}{\theta_*} \frac{\partial \theta}{\partial z} = \phi_h(\zeta) \quad (2.120)$$

$$\frac{k(z-d)}{q_*} \frac{\partial q}{\partial z} = \phi_w(\zeta) \quad (2.121)$$

where z is height in the surface layer (m), d is the displacement height (m), L is the Monin-Obukhov length scale (m) that accounts for buoyancy effects resulting from vertical density gradients (i.e., the atmospheric stability), k is the von Karman constant (Table 2.7), and $|u|$ is the atmospheric wind speed (m s^{-1}). ϕ_m , ϕ_h , and ϕ_w are universal (over any surface) similarity functions of ζ that relate the constant fluxes of momentum, sensible heat, and latent heat to the

mean profile gradients of $|u|$, θ , and q in the surface layer. In neutral conditions, $\phi_m = \phi_h = \phi_w = 1$. The velocity (i.e., friction velocity) u (m s^{-1}), temperature θ (K), and moisture q (kg kg^{-1}) scales are

$$u_*^2 = \sqrt{(\bar{u}'w')^2 + (\bar{v}'w')^2} = \frac{|\tau|}{\rho_{atm}} \quad (2.122)$$

$$\theta_* u_* = -\bar{\theta}'w' = -\frac{H}{\rho_{atm} C_p} \quad (2.123)$$

$$q_* u_* = -\bar{q}'w' = -\frac{E}{\rho_{atm}} \quad (2.124)$$

where $|\tau|$ is the shearing stress ($\text{kg m}^{-1} \text{s}^{-2}$), with zonal and meridional components $\bar{u}'w' = -\frac{\tau_x}{\rho_{atm}}$ and $\bar{v}'w' = -\frac{\tau_y}{\rho_{atm}}$, respectively, H is the sensible heat flux (W m^{-2}) and E is the water vapor flux ($\text{kg m}^{-2} \text{s}^{-1}$).

The length scale L is the Monin-Obukhov length defined as

$$L = -\frac{u_*^3}{k \left(\frac{g}{\theta_{v, atm}} \right) \theta'_v w'} = \frac{u_*^2 \bar{\theta}_{v, atm}}{kg \theta_{v*}} \quad (2.125)$$

where g is the acceleration of gravity (m s^{-2}) (Table 2.7), and $\bar{\theta}_{v, atm} = \bar{\theta}_{atm} (1 + 0.61q_{atm})$ is the reference virtual potential temperature. $L > 0$ indicates stable conditions. $L < 0$ indicates unstable conditions. $L = \infty$ for neutral conditions. The temperature scale θ_{v*} is defined as

$$\theta_{v*} u_* = [\theta_* (1 + 0.61q_{atm}) + 0.61\bar{\theta}_{atm} q_*] u_* \quad (2.126)$$

where $\bar{\theta}_{atm}$ is the atmospheric potential temperature.

Following [Panofsky and Dutton \(1984\)](#), the differential equations for $\phi_m(\zeta)$, $\phi_h(\zeta)$, and $\phi_w(\zeta)$ can be integrated formally without commitment to their exact forms. Integration between two arbitrary heights in the surface layer z_2 and z_1 ($z_2 > z_1$) with horizontal winds $|u|_1$ and $|u|_2$, potential temperatures θ_1 and θ_2 , and specific humidities q_1 and q_2 results in

$$|u|_2 - |u|_1 = \frac{u_*}{k} \left[\ln \left(\frac{z_2 - d}{z_1 - d} \right) - \psi_m \left(\frac{z_2 - d}{L} \right) + \psi_m \left(\frac{z_1 - d}{L} \right) \right] \quad (2.127)$$

$$\theta_2 - \theta_1 = \frac{\theta_*}{k} \left[\ln \left(\frac{z_2 - d}{z_1 - d} \right) - \psi_h \left(\frac{z_2 - d}{L} \right) + \psi_h \left(\frac{z_1 - d}{L} \right) \right] \quad (2.128)$$

$$q_2 - q_1 = \frac{q_*}{k} \left[\ln \left(\frac{z_2 - d}{z_1 - d} \right) - \psi_w \left(\frac{z_2 - d}{L} \right) + \psi_w \left(\frac{z_1 - d}{L} \right) \right]. \quad (2.129)$$

The functions $\psi_m(\zeta)$, $\psi_h(\zeta)$, and $\psi_w(\zeta)$ are defined as

$$\psi_m(\zeta) = \int_{z_{0m}/L}^{\zeta} \frac{[1 - \phi_m(x)]}{x} dx \quad (2.130)$$

$$\psi_h(\zeta) = \int_{z_{0h}/L}^{\zeta} \frac{[1 - \phi_h(x)]}{x} dx \quad (2.131)$$

$$\psi_w(\zeta) = \int_{z_{0w}/L}^{\zeta} \frac{[1 - \phi_w(x)]}{x} dx \quad (2.132)$$

where z_{0m} , z_{0h} , and z_{0w} are the roughness lengths (m) for momentum, sensible heat, and water vapor, respectively.

Defining the surface values

$$|u|_1 = 0 \text{ at } z_1 = z_{0m} + d,$$

$$\theta_1 = \theta_s \text{ at } z_1 = z_{0h} + d, \text{ and}$$

$$q_1 = q_s \text{ at } z_1 = z_{0w} + d,$$

and the atmospheric values at $z_2 = z_{atm,x}$

$$|u|_2 = V_a = \sqrt{u_{atm}^2 + v_{atm}^2 + U_c^2} \geq 1, \quad (2.133)$$

$$\theta_2 = \theta_{atm}, \text{ and}$$

$$q_2 = q_{atm},$$

the integral forms of the flux-gradient relations are

$$V_a = \frac{u_*}{k} \left[\ln \left(\frac{z_{atm,m} - d}{z_{0m}} \right) - \psi_m \left(\frac{z_{atm,m} - d}{L} \right) + \psi_m \left(\frac{z_{0m}}{L} \right) \right] \quad (2.134)$$

$$\theta_{atm} - \theta_s = \frac{\theta_*}{k} \left[\ln \left(\frac{z_{atm,h} - d}{z_{0h}} \right) - \psi_h \left(\frac{z_{atm,h} - d}{L} \right) + \psi_h \left(\frac{z_{0h}}{L} \right) \right] \quad (2.135)$$

$$q_{atm} - q_s = \frac{q_*}{k} \left[\ln \left(\frac{z_{atm,w} - d}{z_{0w}} \right) - \psi_w \left(\frac{z_{atm,w} - d}{L} \right) + \psi_w \left(\frac{z_{0w}}{L} \right) \right]. \quad (2.136)$$

The constraint $V_a \geq 1$ is required simply for numerical reasons to prevent H and E from becoming small with small wind speeds. The convective velocity U_c accounts for the contribution of large eddies in the convective boundary layer to surface fluxes as follows

$$U_c = \begin{cases} 0 & \zeta \geq 0 \text{ (stable)} \\ \beta w_* & \zeta < 0 \text{ (unstable)} \end{cases} \quad (2.137)$$

where w_* is the convective velocity scale

$$w_* = \left(\frac{-gu\theta_{v*}z_i}{\theta_{v,atm}} \right)^{1/3}, \quad (2.138)$$

$z_i = 1000$ is the convective boundary layer height (m), and $\beta = 1$.

The momentum flux gradient relations are (Zeng et al. 1998)

$$\begin{aligned} \phi_m(\zeta) &= 0.7k^{2/3}(-\zeta)^{1/3} & \text{for } \zeta < -1.574 & \text{(very unstable)} \\ \phi_m(\zeta) &= (1 - 16\zeta)^{-1/4} & \text{for } -1.574 \leq \zeta < 0 & \text{(unstable)} \\ \phi_m(\zeta) &= 1 + 5\zeta & \text{for } 0 \leq \zeta \leq 1 & \text{(stable)} \\ \phi_m(\zeta) &= 5 + \zeta & \text{for } \zeta > 1 & \text{(very stable).} \end{aligned} \quad (2.139)$$

The sensible and latent heat flux gradient relations are (Zeng et al. 1998)

$$\begin{aligned}\phi_h(\zeta) &= \phi_w(\zeta) = 0.9k^{4/3}(-\zeta)^{-1/3} & \text{for } \zeta < -0.465 & \text{(very unstable)} \\ \phi_h(\zeta) &= \phi_w(\zeta) = (1 - 16\zeta)^{-1/2} & \text{for } -0.465 \leq \zeta < 0 & \text{(unstable)} \\ \phi_h(\zeta) &= \phi_w(\zeta) = 1 + 5\zeta & \text{for } 0 \leq \zeta \leq 1 & \text{(stable)} \\ \phi_h(\zeta) &= \phi_w(\zeta) = 5 + \zeta & \text{for } \zeta > 1 & \text{(very stable).}\end{aligned}\quad (2.140)$$

To ensure continuous functions of $\phi_m(\zeta)$, $\phi_h(\zeta)$, and $\phi_w(\zeta)$, the simplest approach (i.e., without considering any transition regimes) is to match the relations for very unstable and unstable conditions at $\zeta_m = -1.574$ for $\phi_m(\zeta)$ and $\zeta_h = \zeta_w = -0.465$ for $\phi_h(\zeta) = \phi_w(\zeta)$ (Zeng et al. 1998). The flux gradient relations can be integrated to yield wind profiles for the following conditions:

Very unstable ($\zeta < -1.574$)

$$V_a = \frac{u_*}{k} \left\{ \left[\ln \frac{\zeta_m L}{z_{0m}} - \psi_m(\zeta_m) \right] + 1.14 \left[(-\zeta)^{1/3} - (-\zeta_m)^{1/3} \right] + \psi_m \left(\frac{z_{0m}}{L} \right) \right\} \quad (2.141)$$

Unstable ($-1.574 \leq \zeta < 0$)

$$V_a = \frac{u_*}{k} \left\{ \left[\ln \frac{z_{atm,m} - d}{z_{0m}} - \psi_m(\zeta) \right] + \psi_m \left(\frac{z_{0m}}{L} \right) \right\} \quad (2.142)$$

Stable ($0 \leq \zeta \leq 1$)

$$V_a = \frac{u_*}{k} \left\{ \left[\ln \frac{z_{atm,m} - d}{z_{0m}} + 5\zeta \right] - 5 \frac{z_{0m}}{L} \right\} \quad (2.143)$$

Very stable ($\zeta > 1$)

$$V_a = \frac{u_*}{k} \left\{ \left[\ln \frac{L}{z_{0m}} + 5 \right] + [5 \ln \zeta + \zeta - 1] - 5 \frac{z_{0m}}{L} \right\} \quad (2.144)$$

where

$$\psi_m(\zeta) = 2 \ln \left(\frac{1+x}{2} \right) + \ln \left(\frac{1+x^2}{2} \right) - 2 \tan^{-1} x + \frac{\pi}{2} \quad (2.145)$$

and

$$x = (1 - 16\zeta)^{1/4}.$$

The potential temperature profiles are:

Very unstable ($\zeta < -0.465$)

$$\theta_{atm} - \theta_s = \frac{\theta_*}{k} \left\{ \left[\ln \frac{\zeta_h L}{z_{0h}} - \psi_h(\zeta_h) \right] + 0.8 \left[(-\zeta_h)^{-1/3} - (-\zeta)^{-1/3} \right] + \psi_h \left(\frac{z_{0h}}{L} \right) \right\} \quad (2.146)$$

Unstable ($-0.465 \leq \zeta < 0$)

$$\theta_{atm} - \theta_s = \frac{\theta_*}{k} \left\{ \left[\ln \frac{z_{atm,h} - d}{z_{0h}} - \psi_h(\zeta) \right] + \psi_h \left(\frac{z_{0h}}{L} \right) \right\} \quad (2.147)$$

Stable ($0 \leq \zeta \leq 1$)

$$\theta_{atm} - \theta_s = \frac{\theta_*}{k} \left\{ \left[\ln \frac{z_{atm,h} - d}{z_{0h}} + 5\zeta \right] - 5 \frac{z_{0h}}{L} \right\} \quad (2.148)$$

Very stable ($\zeta > 1$)

$$\theta_{atm} - \theta_s = \frac{\theta_*}{k} \left\{ \left[\ln \frac{L}{z_{0h}} + 5 \right] + [5 \ln \zeta + \zeta - 1] - 5 \frac{z_{0h}}{L} \right\}. \quad (2.149)$$

The specific humidity profiles are:

Very unstable ($\zeta < -0.465$)

$$q_{atm} - q_s = \frac{q_*}{k} \left\{ \left[\ln \frac{\zeta_w L}{z_{0w}} - \psi_w(\zeta_w) \right] + 0.8 \left[(-\zeta_w)^{-1/3} - (-\zeta)^{-1/3} \right] + \psi_w \left(\frac{z_{0w}}{L} \right) \right\} \quad (2.150)$$

Unstable ($-0.465 \leq \zeta < 0$)

$$q_{atm} - q_s = \frac{q_*}{k} \left\{ \left[\ln \frac{z_{atm,w} - d}{z_{0w}} - \psi_w(\zeta) \right] + \psi_w \left(\frac{z_{0w}}{L} \right) \right\} \quad (2.151)$$

Stable ($0 \leq \zeta \leq 1$)

$$q_{atm} - q_s = \frac{q_*}{k} \left\{ \left[\ln \frac{z_{atm,w} - d}{z_{0w}} + 5\zeta \right] - 5 \frac{z_{0w}}{L} \right\} \quad (2.152)$$

Very stable ($\zeta > 1$)

$$q_{atm} - q_s = \frac{q_*}{k} \left\{ \left[\ln \frac{L}{z_{0w}} + 5 \right] + [5 \ln \zeta + \zeta - 1] - 5 \frac{z_{0w}}{L} \right\} \quad (2.153)$$

where

$$\psi_h(\zeta) = \psi_w(\zeta) = 2 \ln \left(\frac{1+x^2}{2} \right). \quad (2.154)$$

Using the definitions of u_* , θ_* , and q_* , an iterative solution of these equations can be used to calculate the surface momentum, sensible heat, and water vapor flux using atmospheric and surface values for $|u|$, θ , and q except that L depends on u_* , θ_* , and q_* . However, the bulk Richardson number

$$R_{iB} = \frac{\theta_{v,atm} - \theta_{v,s}}{\overline{\theta}_{v,atm}} \frac{g(z_{atm,m} - d)}{V_a^2} \quad (2.155)$$

is related to ζ (Arya 2001) as

$$R_{iB} = \zeta \left[\ln \left(\frac{z_{atm,h} - d}{z_{0h}} \right) - \psi_h(\zeta) \right] \left[\ln \left(\frac{z_{atm,m} - d}{z_{0m}} \right) - \psi_m(\zeta) \right]^{-2}. \quad (2.156)$$

Using $\phi_h = \phi_m^2 = (1 - 16\zeta)^{-1/2}$ for unstable conditions and $\phi_h = \phi_m = 1 + 5\zeta$ for stable conditions to determine $\psi_m(\zeta)$ and $\psi_h(\zeta)$, the inverse relationship $\zeta = f(R_{iB})$ can be solved to obtain a first guess for ζ and thus L from

$$\begin{aligned} \zeta &= \frac{R_{iB} \ln \left(\frac{z_{atm,m} - d}{z_{0m}} \right)}{1 - 5 \min(R_{iB}, 0.19)} & 0.01 \leq \zeta \leq 2 & \text{for } R_{iB} \geq 0 \text{ (neutral or stable)} \\ \zeta &= R_{iB} \ln \left(\frac{z_{atm,m} - d}{z_{0m}} \right) & -100 \leq \zeta \leq -0.01 & \text{for } R_{iB} < 0 \text{ (unstable)} \end{aligned} \quad (2.157)$$

Upon iteration (section 2.5.3), the following is used to determine ζ and thus L

$$\zeta = \frac{(z_{atm,m} - d) kg\theta_{v*}}{u_*^2 \overline{\theta}_{v,atm}} \quad (2.158)$$

where

$$\begin{aligned} 0.01 \leq \zeta \leq 2 & \quad \text{for } \zeta \geq 0 \text{ (neutral or stable)} \\ -100 \leq \zeta \leq -0.01 & \quad \text{for } \zeta < 0 \text{ (unstable)} \end{aligned}.$$

The difference in virtual potential air temperature between the reference height and the surface is

$$\theta_{v,atm} - \theta_{v,s} = (\theta_{atm} - \theta_s) (1 + 0.61q_{atm}) + 0.61\overline{\theta}_{atm} (q_{atm} - q_s). \quad (2.159)$$

The momentum, sensible heat, and water vapor fluxes between the surface and the atmosphere can also be written in the form

$$\tau_x = -\rho_{atm} \frac{(u_{atm} - u_s)}{r_{am}} \quad (2.160)$$

$$\tau_y = -\rho_{atm} \frac{(v_{atm} - v_s)}{r_{am}} \quad (2.161)$$

$$H = -\rho_{atm} C_p \frac{(\theta_{atm} - \theta_s)}{r_{ah}} \quad (2.162)$$

$$E = -\rho_{atm} \frac{(q_{atm} - q_s)}{r_{aw}} \quad (2.163)$$

where the aerodynamic resistances (s m^{-1}) are

$$r_{am} = \frac{V_a}{u_*^2} = \frac{1}{k^2 V_a} \left[\ln \left(\frac{z_{atm,m} - d}{z_{0m}} \right) - \psi_m \left(\frac{z_{atm,m} - d}{L} \right) + \psi_m \left(\frac{z_{0m}}{L} \right) \right]^2 \quad (2.164)$$

$$r_{ah} = \frac{\theta_{atm} - \theta_s}{\theta_* u_*} = \frac{1}{k^2 V_a} \left[\ln \left(\frac{z_{atm,m} - d}{z_{0m}} \right) - \psi_m \left(\frac{z_{atm,m} - d}{L} \right) + \psi_m \left(\frac{z_{0m}}{L} \right) \right] \left[\ln \left(\frac{z_{atm,h} - d}{z_{0h}} \right) - \psi_h \left(\frac{z_{atm,h} - d}{L} \right) + \psi_h \left(\frac{z_{0h}}{L} \right) \right] \quad (2.165)$$

$$r_{aw} = \frac{q_{atm} - q_s}{q_* u_*} = \frac{1}{k^2 V_a} \left[\ln \left(\frac{z_{atm,m} - d}{z_{0m}} \right) - \psi_m \left(\frac{z_{atm,m} - d}{L} \right) + \psi_m \left(\frac{z_{0m}}{L} \right) \right] \left[\ln \left(\frac{z_{atm,w} - d}{z_{0w}} \right) - \psi_w \left(\frac{z_{atm,w} - d}{L} \right) + \psi_w \left(\frac{z_{0w}}{L} \right) \right]. \quad (2.166)$$

A 2-m height “screen” temperature is useful for comparison with observations

$$T_{2m} = \theta_s + \frac{\theta_*}{k} \left[\ln \left(\frac{2 + z_{0h}}{z_{0h}} \right) - \psi_h \left(\frac{2 + z_{0h}}{L} \right) + \psi_h \left(\frac{z_{0h}}{L} \right) \right] \quad (2.167)$$

where for convenience, “2-m” is defined as 2 m above the apparent sink for sensible heat ($z_{0h} + d$). Similarly, a 2-m height specific humidity is defined as

$$q_{2m} = q_s + \frac{q_*}{k} \left[\ln \left(\frac{2 + z_{0w}}{z_{0w}} \right) - \psi_w \left(\frac{2 + z_{0w}}{L} \right) + \psi_w \left(\frac{z_{0w}}{L} \right) \right]. \quad (2.168)$$

Relative humidity is

$$RH_{2m} = \min \left(100, \frac{q_{2m}}{q_{sat}^{T_{2m}}} \times 100 \right) \quad (2.169)$$

where $q_{sat}^{T_{2m}}$ is the saturated specific humidity at the 2-m temperature T_{2m} (section 2.5.5).

A 10-m wind speed is calculated as (note that this is not consistent with the 10-m wind speed calculated for the dust model as described in Chapter 2.30)

$$u_{10m} = \begin{cases} V_a & z_{atm,m} \leq 10 \\ V_a - \frac{u_*}{k} \left[\ln \left(\frac{z_{atm,m} - d}{10 + z_{0m}} \right) - \psi_m \left(\frac{z_{atm,m} - d}{L} \right) + \psi_m \left(\frac{10 + z_{0m}}{L} \right) \right] & z_{atm,m} > 10 \end{cases} \quad (2.170)$$

2.5.2 Sensible and Latent Heat Fluxes for Non-Vegetated Surfaces

Surfaces are considered non-vegetated for the surface flux calculations if leaf plus stem area index $L + S < 0.05$ (section 2.2.1). By definition, this includes bare soil, wetlands, and glaciers. The solution for lakes is described in Chapter 2.12. For these surfaces, the surface may be exposed to the atmosphere, snow covered, and/or surface water covered, so that the sensible heat flux H_g (W m^{-2}) is, with reference to Figure 2.4,

$$H_g = (1 - f_{sno} - f_{h2osfc}) H_{soil} + f_{sno} H_{snow} + f_{h2osfc} H_{h2osfc} \quad (2.171)$$

where $(1 - f_{sno} - f_{h2osfc})$, f_{sno} , and f_{h2osfc} are the exposed, snow covered, and surface water covered fractions of the grid cell. The individual fluxes based on the temperatures of the soil T_1 , snow T_{snl+1} , and surface water T_{h2osfc} are

$$H_{soil} = -\rho_{atm} C_p \frac{(\theta_{atm} - T_1)}{r_{ah}} \quad (2.172)$$

$$H_{sno} = -\rho_{atm} C_p \frac{(\theta_{atm} - T_{snl+1})}{r_{ah}} \quad (2.173)$$

$$H_{h2osfc} = -\rho_{atm} C_p \frac{(\theta_{atm} - T_{h2osfc})}{r_{ah}} \quad (2.174)$$

where ρ_{atm} is the density of atmospheric air (kg m^{-3}), C_p is the specific heat capacity of air ($\text{J kg}^{-1} \text{K}^{-1}$) (Table 2.7), θ_{atm} is the atmospheric potential temperature (K), and r_{ah} is the aerodynamic resistance to sensible heat transfer (s m^{-1}).

The water vapor flux E_g ($\text{kg m}^{-2} \text{s}^{-1}$) is, with reference to Figure 2.5,

$$E_g = (1 - f_{sno} - f_{h2osfc}) E_{soil} + f_{sno} E_{snow} + f_{h2osfc} E_{h2osfc} \quad (2.175)$$

$$E_{soil} = -\frac{\rho_{atm} (q_{atm} - q_{soil})}{r_{aw} + r_{soil}} \quad (2.176)$$

$$E_{sno} = -\frac{\rho_{atm} (q_{atm} - q_{sno})}{r_{aw}} \quad (2.177)$$

$$E_{h2osfc} = -\frac{\rho_{atm} (q_{atm} - q_{h2osfc})}{r_{aw}} \quad (2.178)$$

where q_{atm} is the atmospheric specific humidity (kg kg^{-1}), q_{soil} , q_{sno} , and q_{h2osfc} are the specific humidities (kg kg^{-1}) of the soil, snow, and surface water, respectively, r_{aw} is the aerodynamic resistance to water vapor transfer (s m^{-1}), and r_{soil} is the soil resistance to water vapor transfer (s m^{-1}). The specific humidities of the snow q_{sno} and surface water q_{h2osfc} are assumed to be at the saturation specific humidity of their respective temperatures

$$q_{sno} = q_{sat}^{T_{snl+1}} \quad (2.179)$$

$$q_{h2osfc} = q_{sat}^{T_{h2osfc}} \quad (2.180)$$

The specific humidity of the soil surface q_{soil} is assumed to be proportional to the saturation specific humidity

$$q_{soil} = \alpha_{soil} q_{sat}^{T_1} \quad (2.181)$$

where $q_{sat}^{T_1}$ is the saturated specific humidity at the soil surface temperature T_1 (section 2.5.5). The factor α_{soil} is a function of the surface soil water matric potential ψ as in [Philip \(1957\)](#)

$$\alpha_{soil} = \exp\left(\frac{\psi_1 g}{1 \times 10^3 R_{wv} T_1}\right) \quad (2.182)$$

where R_{wv} is the gas constant for water vapor ($\text{J kg}^{-1} \text{K}^{-1}$) ([Table 2.7](#)), g is the gravitational acceleration (m s^{-2}) ([Table 2.7](#)), and ψ_1 is the soil water matric potential of the top soil layer (mm). The soil water matric potential ψ_1 is

$$\psi_1 = \psi_{sat, 1} s_1^{-B_1} \geq -1 \times 10^8 \quad (2.183)$$

where $\psi_{sat, 1}$ is the saturated matric potential (mm) (section 2.7.3), B_1 is the [Clapp and Hornberger \(1978\)](#) parameter (section 2.7.3), and s_1 is the wetness of the top soil layer with respect to saturation. The surface wetness s_1 is a function of the liquid water and ice content

$$s_1 = \frac{1}{\Delta z_1 \theta_{sat, 1}} \left[\frac{w_{liq, 1}}{\rho_{liq}} + \frac{w_{ice, 1}}{\rho_{ice}} \right] \quad 0.01 \leq s_1 \leq 1.0 \quad (2.184)$$

where Δz_1 is the thickness of the top soil layer (m), ρ_{liq} and ρ_{ice} are the density of liquid water and ice (kg m^{-3}) ([Table 2.7](#)), $w_{liq, 1}$ and $w_{ice, 1}$ are the mass of liquid water and ice of the top soil layer (kg m^{-2}) ([Chapter 2.7](#)), and $\theta_{sat, 1}$ is the saturated volumetric water content (i.e., porosity) of the top soil layer ($\text{mm}^3 \text{mm}^{-3}$) (section 2.7.3). If $q_{sat}^{T_1} > q_{atm}$ and $q_{atm} > q_{soil}$, then $q_{soil} = q_{atm}$ and $\frac{dq_{soil}}{dT} = 0$. This prevents large increases (decreases) in q_{soil} for small increases (decreases) in soil moisture in very dry soils.

The resistance to water vapor transfer occurring within the soil matrix r_{soil} (s m^{-1}) is

$$r_{soil} = \frac{DSL}{D_v \tau} \quad (2.185)$$

where DSL is the thickness of the dry surface layer (m), D_v is the molecular diffusivity of water vapor in air ($\text{m}^2 \text{s}^{-2}$) and τ (*unitless*) describes the tortuosity of the vapor flow paths through the soil matrix ([Swenson and Lawrence 2014](#)).

The thickness of the dry surface layer is given by

$$DSL = \begin{cases} D_{max} \frac{(\theta_{init} - \theta_1)}{(\theta_{init} - \theta_{air})} & \theta_1 < \theta_{init} \\ 0 & \theta_1 \geq \theta_{init} \end{cases} \quad (2.186)$$

where D_{max} is a parameter specifying the length scale of the maximum DSL thickness (default value = 15 mm), θ_{init} ($\text{mm}^3 \text{mm}^{-3}$) is the moisture value at which the DSL initiates, θ_1 ($\text{mm}^3 \text{mm}^{-3}$) is the moisture value of the top model soil layer, and θ_{air} ($\text{mm}^3 \text{mm}^{-3}$) is the ‘air dry’ soil moisture value ([Dingman 2002](#)):

$$\theta_{air} = \Phi \left(\frac{\Psi_{sat}}{\Psi_{air}} \right)^{\frac{1}{B_1}}. \quad (2.187)$$

where Φ is the porosity ($\text{mm}^3 \text{mm}^{-3}$), Ψ_{sat} is the saturated soil matric potential (mm), $\Psi_{air} = 10^7$ mm is the air dry matric potential, and B_1 is a function of soil texture (section 2.7.3).

The soil tortuosity is

$$\tau = \Phi_{air}^2 \left(\frac{\Phi_{air}}{\Phi} \right)^{\frac{3}{B_1}} \quad (2.188)$$

where Φ_{air} ($\text{mm}^3 \text{mm}^{-3}$) is the air filled pore space

$$\Phi_{air} = \Phi - \theta_{air}. \quad (2.189)$$

D_v depends on temperature

$$D_v = 2.12 \times 10^{-5} \left(\frac{T_1}{T_f} \right)^{1.75}. \quad (2.190)$$

where T_1 (K) is the temperature of the top soil layer and T_f (K) is the freezing temperature of water (Table 2.7).

The roughness lengths used to calculate r_{am} , r_{ah} , and r_{aw} are $z_{0m} = z_{0m,g}$, $z_{0h} = z_{0h,g}$, and $z_{0w} = z_{0w,g}$. The displacement height $d = 0$. The momentum roughness length is $z_{0m,g} = 0.01$ for soil, glaciers, and wetland, and $z_{0m,g} = 0.0024$ for snow-covered surfaces ($f_{sno} > 0$). In general, z_{0m} is different from z_{0h} because the transfer of momentum is affected by pressure fluctuations in the turbulent waves behind the roughness elements, while for heat and water vapor transfer no such dynamical mechanism exists. Rather, heat and water vapor must be transferred by molecular diffusion across the interfacial sublayer. The following relation from [Zilitinkevich \(1970\)](#) is adopted by [Zeng and Dickinson 1998](#)

$$z_{0h,g} = z_{0w,g} = z_{0m,g} e^{-a(u_* z_{0m,g}/v)^{0.45}} \quad (2.191)$$

where the quantity $uz_{0m,g}/v$ is the roughness Reynolds number (and may be interpreted as the Reynolds number of the smallest turbulent eddy in the flow) with the kinematic viscosity of air $v = 1.5 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ and $a = 0.13$.

The numerical solution for the fluxes of momentum, sensible heat, and water vapor flux from non-vegetated surfaces proceeds as follows:

1. An initial guess for the wind speed V_a is obtained from (2.133) assuming an initial convective velocity $U_c = 0 \text{ m s}^{-1}$ for stable conditions ($\theta_{v,atm} - \theta_{v,s} \geq 0$ as evaluated from (2.159)) and $U_c = 0.5$ for unstable conditions ($\theta_{v,atm} - \theta_{v,s} < 0$).
2. An initial guess for the Monin-Obukhov length L is obtained from the bulk Richardson number using (2.155) and (2.157).
3. The following system of equations is iterated three times:
 4. Friction velocity u_* ((2.141), (2.142), (2.143), (2.144))
 5. Potential temperature scale θ_* ((2.146), (2.147), (2.148), (2.149))
 6. Humidity scale q_* ((2.150), (2.151), (2.152), (2.153))
 7. Roughness lengths for sensible $z_{0h,g}$ and latent heat $z_{0w,g}$ ((2.191))
 8. Virtual potential temperature scale θ_{v*} ((2.126))
 9. Wind speed including the convective velocity, V_a ((2.133))
 10. Monin-Obukhov length L ((2.158))
 11. Aerodynamic resistances r_{am} , r_{ah} , and r_{aw} ((2.164), (2.165), (2.166))
 12. Momentum fluxes τ_x , τ_y ((2.114), (2.115))
 13. Sensible heat flux H_g ((2.171))
 14. Water vapor flux E_g ((2.175))
 15. 2-m height air temperature T_{2m} and specific humidity q_{2m} ((2.167), (2.168))

The partial derivatives of the soil surface fluxes with respect to ground temperature, which are needed for the soil temperature calculations (section 2.6.1) and to update the soil surface fluxes (section 2.5.4), are

$$\frac{\partial H_g}{\partial T_g} = \frac{\rho_{atm} C_p}{r_{ah}} \quad (2.192)$$

$$\frac{\partial E_g}{\partial T_g} = \frac{\beta_{soil} \rho_{atm}}{r_{aw}} \frac{dq_g}{dT_g} \quad (2.193)$$

where

$$\frac{dq_g}{dT_g} = (1 - f_{sno} - f_{h2osfc}) \alpha_{soil} \frac{dq_{sat}^{T_{soil}}}{dT_{soil}} + f_{sno} \frac{dq_{sat}^{T_{sno}}}{dT_{sno}} + f_{h2osfc} \frac{dq_{sat}^{T_{h2osfc}}}{dT_{h2osfc}}. \quad (2.194)$$

The partial derivatives $\frac{\partial r_{ah}}{\partial T_g}$ and $\frac{\partial r_{aw}}{\partial T_g}$, which cannot be determined analytically, are ignored for $\frac{\partial H_g}{\partial T_g}$ and $\frac{\partial E_g}{\partial T_g}$.

2.5.3 Sensible and Latent Heat Fluxes and Temperature for Vegetated Surfaces

In the case of a vegetated surface, the sensible heat H and water vapor flux E are partitioned into vegetation and ground fluxes that depend on vegetation T_v and ground T_g temperatures in addition to surface temperature T_s and specific humidity q_s . Because of the coupling between vegetation temperature and fluxes, Newton-Raphson iteration is used to solve for the vegetation temperature and the sensible heat and water vapor fluxes from vegetation simultaneously using the ground temperature from the previous time step. In section 2.5.3, the equations used in the iteration scheme are derived. Details on the numerical scheme are provided in section 2.5.3.

Theory

The air within the canopy is assumed to have negligible capacity to store heat so that the sensible heat flux H between the surface at height $z_{0h} + d$ and the atmosphere at height $z_{atm,h}$ must be balanced by the sum of the sensible heat from the vegetation H_v and the ground H_g

$$H = H_v + H_g \quad (2.195)$$

where, with reference to Figure 2.4,

$$H = -\rho_{atm} C_p \frac{(\theta_{atm} - T_s)}{r_{ah}} \quad (2.196)$$

$$H_v = -\rho_{atm} C_p (T_s - T_v) \frac{(L + S)}{r_b} \quad (2.197)$$

$$H_g = (1 - f_{sno} - f_{h2osfc}) H_{soil} + f_{sno} H_{snow} + f_{h2osfc} H_{h2osfc}, \quad (2.198)$$

where

$$H_{soil} = -\rho_{atm} C_p \frac{(T_s - T_1)}{r'_{ah}} \quad (2.199)$$

$$H_{snow} = -\rho_{atm} C_p \frac{(T_s - T_{snl+1})}{r'_{ah}} \quad (2.200)$$

$$H_{h2osfc} = -\rho_{atm} C_p \frac{(T_s - T_{h2osfc})}{r'_{ah}} \quad (2.201)$$

where ρ_{atm} is the density of atmospheric air (kg m^{-3}), C_p is the specific heat capacity of air ($\text{J kg}^{-1} \text{ K}^{-1}$) (Table 2.7), θ_{atm} is the atmospheric potential temperature (K), and r_{ah} is the aerodynamic resistance to sensible heat transfer (s m^{-1}).

Here, T_s is the surface temperature at height $z_{0h} + d$, also referred to as the canopy air temperature. L and S are the exposed leaf and stem area indices (section 2.2.1), r_b is the leaf boundary layer resistance (s m^{-1}), and r'_{ah} is the aerodynamic resistance (s m^{-1}) to heat transfer between the ground at height z_{0h} and the canopy air at height $z_{0h} + d$.

Equations (2.195) - (2.198) can be solved for the canopy air temperature T_s

$$T_s = \frac{c_a^h \theta_{atm} + c_g^h T_g + c_v^h T_v}{c_a^h + c_g^h + c_v^h} \quad (2.202)$$

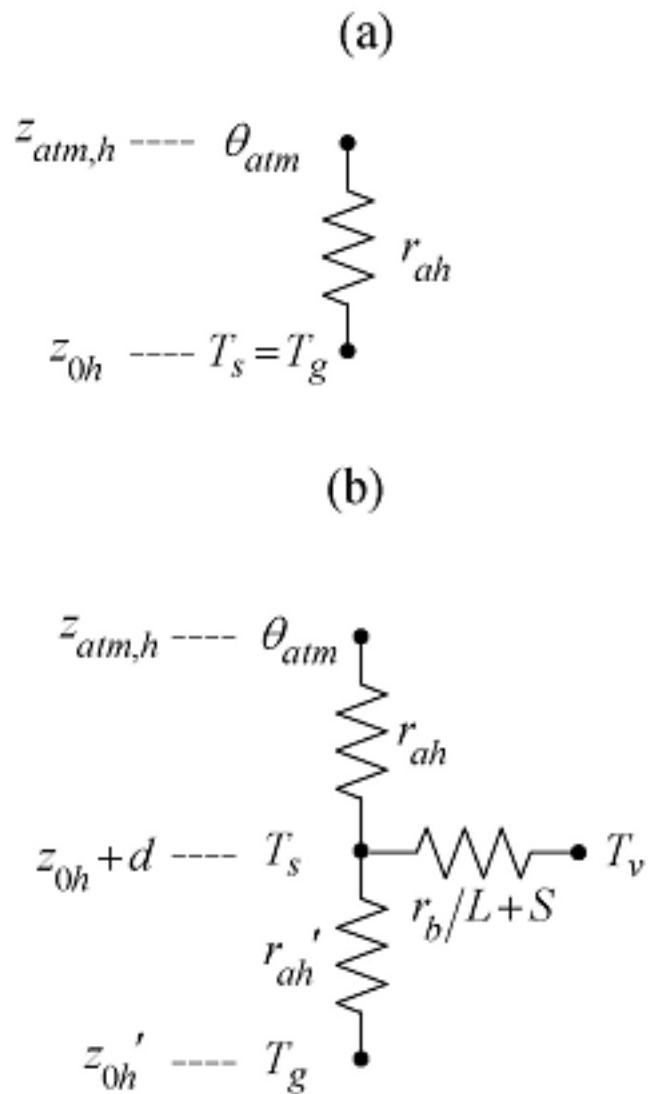


Figure 2.4: Figure Schematic diagram of sensible heat fluxes for (a) non-vegetated surfaces and (b) vegetated surfaces.

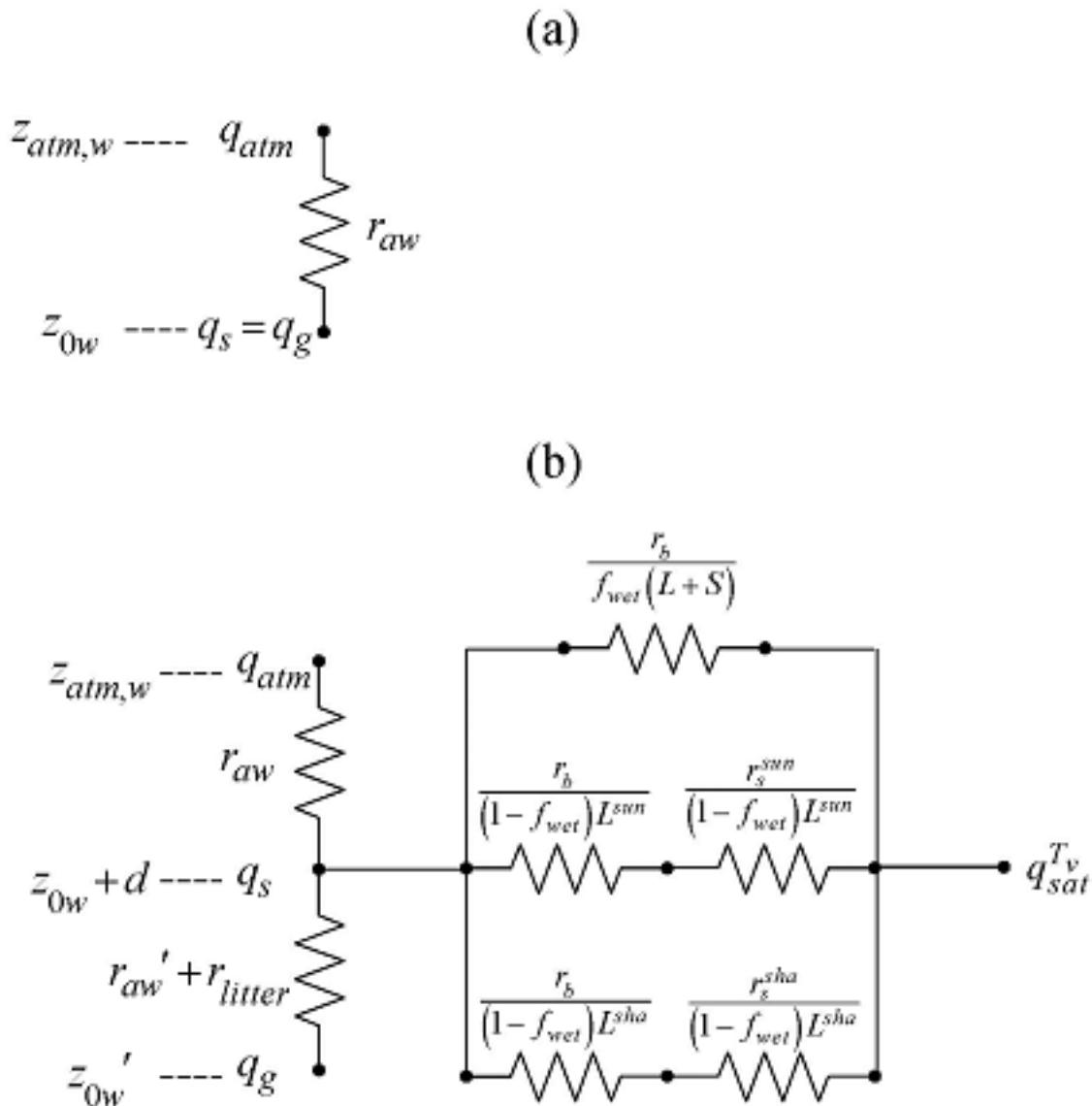


Figure 2.5: Figure Schematic diagram of water vapor fluxes for (a) non-vegetated surfaces and (b) vegetated surfaces.

where

$$c_a^h = \frac{1}{r_{ah}} \quad (2.203)$$

$$c_g^h = \frac{1}{r'_{ah}} \quad (2.204)$$

$$c_v^h = \frac{(L + S)}{r_b} \quad (2.205)$$

are the sensible heat conductances from the canopy air to the atmosphere, the ground to canopy air, and leaf surface to canopy air, respectively (m s^{-1}).

When the expression for T_s is substituted into equation (2.197), the sensible heat flux from vegetation H_v is a function of θ_{atm} , T_g , and T_v

$$H_v = -\rho_{atm} C_p [c_a^h \theta_{atm} + c_g^h T_g - (c_a^h + c_g^h) T_v] \frac{c_v^h}{c_a^h + c_v^h + c_g^h}. \quad (2.206)$$

Similarly, the expression for T_s can be substituted into equation to obtain the sensible heat flux from ground H_g

$$H_g = -\rho_{atm} C_p [c_a^h \theta_{atm} + c_v^h T_v - (c_a^h + c_v^h) T_g] \frac{c_g^h}{c_a^h + c_v^h + c_g^h}. \quad (2.207)$$

The air within the canopy is assumed to have negligible capacity to store water vapor so that the water vapor flux E between the surface at height $z_{0w} + d$ and the atmosphere at height $z_{atm,w}$ must be balanced by the sum of the water vapor flux from the vegetation E_v and the ground E_g

$$E = E_v + E_g \quad (2.208)$$

where, with reference to Figure 2.5,

$$E = -\rho_{atm} \frac{(q_{atm} - q_s)}{r_{aw}} \quad (2.209)$$

$$E_v = -\rho_{atm} \frac{\left(q_s - q_{sat}^{T_v} \right)}{r_{total}} \quad (2.210)$$

$$E_g = (1 - f_{sno} - f_{h2osfc}) E_{soil} + f_{sno} E_{snow} + f_{h2osfc} E_{h2osfc}, \quad (2.211)$$

where

$$E_{soil} = -\rho_{atm} \frac{(q_s - q_{soil})}{r'_{aw} + r_{soil}} \quad (2.212)$$

$$E_{snow} = -\rho_{atm} \frac{(q_s - q_{snow})}{r'_{aw} + r_{soil}} \quad (2.213)$$

$$E_{h2osfc} = -\rho_{atm} \frac{(q_s - q_{h2osfc})}{r'_{aw} + r_{soil}} \quad (2.214)$$

where q_{atm} is the atmospheric specific humidity (kg kg^{-1}), r_{aw} is the aerodynamic resistance to water vapor transfer (s m^{-1}), $q_{sat}^{T_v}$ (kg kg^{-1}) is the saturation water vapor specific humidity at the vegetation temperature (section 2.5.5), q_g , q_{snow} , and q_{h2osfc} are the specific humidities of the soil, snow, and surface water (section 2.5.2), r'_{aw} is the aerodynamic resistance (s m^{-1}) to water vapor transfer between the ground at height z'_{0w} and the canopy air at height $z_{0w} + d$, and r_{soil} ((2.185)) is a resistance to diffusion through the soil (s m^{-1}). r_{total} is the total resistance to water vapor transfer from the canopy to the canopy air and includes contributions from leaf boundary layer and sunlit and shaded stomatal resistances r_b , r_s^{sun} , and r_s^{sha} (Figure 2.5). The water vapor flux from vegetation is the sum of water vapor flux from wetted leaf and stem area E_v^w (evaporation of water intercepted by the canopy) and transpiration from dry leaf surfaces E_v^t

$$E_v = E_v^w + E_v^t. \quad (2.215)$$

Equations (2.208) - (2.211) can be solved for the canopy specific humidity q_s

$$q_s = \frac{c_a^w q_{atm} + c_g^w q_g + c_v^w q_{sat}^{T_v}}{c_a^w + c_v^w + c_g^w} \quad (2.216)$$

where

$$c_a^w = \frac{1}{r_{aw}} \quad (2.217)$$

$$c_v^w = \frac{(L + S)}{r_b} r'' \quad (2.218)$$

$$c_g^w = \frac{1}{r'_{aw} + r_{soil}} \quad (2.219)$$

are the water vapor conductances from the canopy air to the atmosphere, the leaf to canopy air, and ground to canopy air, respectively. The term r'' is determined from contributions by wet leaves and transpiration and limited by available water and potential evaporation as

$$r'' = \begin{cases} \min \left(f_{wet} + r''_{dry}, \frac{E_v^{w,pot} r''_{dry} + \frac{W_{can}}{\Delta t}}{E_v^{w,pot}} \right) & E_v^{w,pot} > 0, \beta_t > 0 \\ \min \left(f_{wet}, \frac{E_v^{w,pot} r''_{dry} + \frac{W_{can}}{\Delta t}}{E_v^{w,pot}} \right) & E_v^{w,pot} > 0, \beta_t \leq 0 \\ 1 & E_v^{w,pot} \leq 0 \end{cases} \quad (2.220)$$

where f_{wet} is the fraction of leaves and stems that are wet (section 2.7.1), W_{can} is canopy water (kg m^{-2}) (section 2.7.1), Δt is the time step (s), and β_t is a soil moisture function limiting transpiration (Chapter 2.9). The potential evaporation from wet foliage per unit wetted area is

$$E_v^{w,pot} = -\frac{\rho_{atm} (q_s - q_{sat}^{T_v})}{r_b}. \quad (2.221)$$

The term r''_{dry} is

$$r''_{dry} = \frac{f_{dry} r_b}{L} \left(\frac{L^{sun}}{r_b + r_s^{sun}} + \frac{L^{sha}}{r_b + r_s^{sha}} \right) \quad (2.222)$$

where f_{dry} is the fraction of leaves that are dry (section 2.7.1), L^{sun} and L^{sha} are the sunlit and shaded leaf area indices (section 2.4.1), and r_s^{sun} and r_s^{sha} are the sunlit and shaded stomatal resistances (s m^{-1}) (Chapter 2.9).

When the expression for q_s is substituted into equation (2.210), the water vapor flux from vegetation E_v is a function of q_{atm} , q_g , and $q_{sat}^{T_v}$

$$E_v = -\rho_{atm} \left[c_a^w q_{atm} + c_g^w q_g - (c_a^w + c_g^w) q_{sat}^{T_v} \right] \frac{c_v^w}{c_a^w + c_v^w + c_g^w}. \quad (2.223)$$

Similarly, the expression for q_s can be substituted into (2.193) to obtain the water vapor flux from the ground beneath the canopy E_g

$$E_g = -\rho_{atm} \left[c_a^w q_{atm} + c_v^w q_{sat}^{T_v} - (c_a^w + c_v^w) q_g \right] \frac{c_g^w}{c_a^w + c_v^w + c_g^w}. \quad (2.224)$$

The aerodynamic resistances to heat (moisture) transfer between the ground at height z'_{0h} (z'_{0w}) and the canopy air at height $z_{0h} + d$ ($z_{0w} + d$) are

$$r'_{ah} = r'_{aw} = \frac{1}{C_s U_{av}} \quad (2.225)$$

where

$$U_{av} = V_a \sqrt{\frac{1}{r_{am} V_a}} = u_* \quad (2.226)$$

is the magnitude of the wind velocity incident on the leaves (equivalent here to friction velocity) (m s^{-1}) and C_s is the turbulent transfer coefficient between the underlying soil and the canopy air. C_s is obtained by interpolation between values for dense canopy and bare soil (Zeng *et al.* 2005)

$$C_s = C_{s, \text{bare}} W + C_{s, \text{dense}} (1 - W) \quad (2.227)$$

where the weight W is

$$W = e^{-(L+S)}. \quad (2.228)$$

The dense canopy turbulent transfer coefficient (Dickinson *et al.* 1993) is

$$C_{s, \text{dense}} = 0.004. \quad (2.229)$$

The bare soil turbulent transfer coefficient is

$$C_{s, \text{bare}} = \frac{k}{a} \left(\frac{z_{0m, g} U_{av}}{v} \right)^{-0.45} \quad (2.230)$$

where the kinematic viscosity of air $v = 1.5 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$ and $a = 0.13$.

The leaf boundary layer resistance r_b is

$$r_b = \frac{1}{C_v} (U_{av}/d_{leaf})^{-1/2} \quad (2.231)$$

where $C_v = 0.01 \text{ ms}^{-1/2}$ is the turbulent transfer coefficient between the canopy surface and canopy air, and d_{leaf} is the characteristic dimension of the leaves in the direction of wind flow (Table 2.17).

The partial derivatives of the fluxes from the soil beneath the canopy with respect to ground temperature, which are needed for the soil temperature calculations (section 2.6.1) and to update the soil surface fluxes (section 2.5.4), are

$$\frac{\partial H_g}{\partial T_g} = \frac{\rho_{atm} C_p}{r'_{ah}} \frac{c_a^h + c_v^h}{c_a^h + c_v^h + c_g^h} \quad (2.232)$$

$$\frac{\partial E_g}{\partial T_g} = \frac{\rho_{atm}}{r'_{aw} + r_{soil}} \frac{c_a^w + c_v^w}{c_a^w + c_v^w + c_g^w} \frac{dq_g}{dT_g}. \quad (2.233)$$

The partial derivatives $\frac{\partial r'_{ah}}{\partial T_g}$ and $\frac{\partial r'_{aw}}{\partial T_g}$, which cannot be determined analytically, are ignored for $\frac{\partial H_g}{\partial T_g}$ and $\frac{\partial E_g}{\partial T_g}$.

The roughness lengths used to calculate r_{am} , r_{ah} , and r_{aw} from (2.164), (2.165), and (2.166) are $z_{0m} = z_{0m,v}$, $z_{0h} = z_{0h,v}$, and $z_{0w} = z_{0w,v}$. The vegetation displacement height d and the roughness lengths are a function of plant height and adjusted for canopy density following [Zeng and Wang \(2007\)](#)

$$z_{0m,v} = z_{0h,v} = z_{0w,v} = \exp [V \ln (z_{top} R_{z0m}) + (1 - V) \ln (z_{0m,g})] \quad (2.234)$$

$$d = z_{top} R_d V \quad (2.235)$$

where z_{top} is canopy top height (m) (Table Prescribed plant functional type heights), R_{z0m} and R_d are the ratio of momentum roughness length and displacement height to canopy top height, respectively ([Table 2.16](#)), and $z_{0m,g}$ is the ground momentum roughness length (m) (section 2.5.2). The fractional weight V is determined from

$$V = \frac{1 - \exp \{-\beta \min [L + S, (L + S)_{cr}]\}}{1 - \exp [-\beta (L + S)_{cr}]} \quad (2.236)$$

where $\beta = 1$ and $(L + S)_{cr} = 2$ ($\text{m}^2 \text{ m}^{-2}$) is a critical value of exposed leaf plus stem area for which z_{0m} reaches its maximum.

Table 2.16: Plant functional type aerodynamic parameters

Plant functional type	R_{z0m}	R_d	d_{leaf} (m)
NET Temperate	0.055	0.67	0.04
NET Boreal	0.055	0.67	0.04
NDT Boreal	0.055	0.67	0.04
BET Tropical	0.075	0.67	0.04
BET temperate	0.075	0.67	0.04
BDT tropical	0.055	0.67	0.04
BDT temperate	0.055	0.67	0.04
BDT boreal	0.055	0.67	0.04
BES temperate	0.120	0.68	0.04
BDS temperate	0.120	0.68	0.04
BDS boreal	0.120	0.68	0.04
C ₃ arctic grass	0.120	0.68	0.04
C ₃ grass	0.120	0.68	0.04
C ₄ grass	0.120	0.68	0.04
Crop R	0.120	0.68	0.04
Crop I	0.120	0.68	0.04
Corn R	0.120	0.68	0.04
Corn I	0.120	0.68	0.04
Temp Cereal R	0.120	0.68	0.04
Temp Cereal I	0.120	0.68	0.04
Winter Cereal R	0.120	0.68	0.04
Winter Cereal I	0.120	0.68	0.04
Soybean R	0.120	0.68	0.04
Soybean I	0.120	0.68	0.04

Numerical Implementation

Canopy energy conservation gives

$$-\vec{S}_v + \vec{L}_v(T_v) + H_v(T_v) + \lambda E_v(T_v) = 0 \quad (2.237)$$

where \vec{S}_v is the solar radiation absorbed by the vegetation (section 2.4.1), \vec{L}_v is the net longwave radiation absorbed by vegetation (section 2.4.2), and H_v and λE_v are the sensible and latent heat fluxes from vegetation, respectively. The term λ is taken to be the latent heat of vaporization λ_{vap} (Table 2.7).

\vec{L}_v , H_v , and λE_v depend on the vegetation temperature T_v . The Newton-Raphson method for finding roots of non-linear systems of equations can be applied to iteratively solve for T_v as

$$\Delta T_v = \frac{\vec{S}_v - \vec{L}_v - H_v - \lambda E_v}{\frac{\partial \vec{L}_v}{\partial T_v} + \frac{\partial H_v}{\partial T_v} + \frac{\partial \lambda E_v}{\partial T_v}} \quad (2.238)$$

where $\Delta T_v = T_v^{n+1} - T_v^n$ and the subscript “n” indicates the iteration.

The partial derivatives are

$$\frac{\partial \vec{L}_v}{\partial T_v} = 4\varepsilon_v \sigma [2 - \varepsilon_v (1 - \varepsilon_g)] T_v^3 \quad (2.239)$$

$$\frac{\partial H_v}{\partial T_v} = \rho_{atm} C_p (c_a^h + c_g^h) \frac{c_v^h}{c_a^h + c_v^h + c_g^h} \quad (2.240)$$

$$\frac{\partial \lambda E_v}{\partial T_v} = \lambda \rho_{atm} (c_a^w + c_g^w) \frac{c_v^w}{c_a^w + c_v^w + c_g^w} \frac{dq_{sat}^{T_v}}{dT_v}. \quad (2.241)$$

The partial derivatives $\frac{\partial r_{ah}}{\partial T_v}$ and $\frac{\partial r_{aw}}{\partial T_v}$, which cannot be determined analytically, are ignored for $\frac{\partial H_v}{\partial T_v}$ and $\frac{\partial \lambda E_v}{\partial T_v}$. However, if ζ changes sign more than four times during the temperature iteration, $\zeta = -0.01$. This helps prevent “flip-flopping” between stable and unstable conditions. The total water vapor flux E_v , transpiration flux E_v^t , and sensible heat flux H_v are updated for changes in leaf temperature as

$$E_v = -\rho_{atm} \left[c_a^w q_{atm} + c_g^w q_g - (c_a^w + c_g^w) \left(q_{sat}^{T_v} + \frac{dq_{sat}^{T_v}}{dT_v} \Delta T_v \right) \right] \frac{c_v^w}{c_a^w + c_v^w + c_g^w} \quad (2.242)$$

$$E_v^t = -r_{dry}'' \rho_{atm} \left[c_a^w q_{atm} + c_g^w q_g - (c_a^w + c_g^w) \left(q_{sat}^{T_v} + \frac{dq_{sat}^{T_v}}{dT_v} \Delta T_v \right) \right] \frac{c_v^h}{c_a^w + c_v^w + c_g^w} \quad (2.243)$$

$$H_v = -\rho_{atm} C_p [c_a^h \theta_{atm} + c_g^h T_g - (c_a^h + c_g^h) (T_v + \Delta T_v)] \frac{c_v^h}{c_a^h + c_v^h + c_g^h}. \quad (2.244)$$

The numerical solution for vegetation temperature and the fluxes of momentum, sensible heat, and water vapor flux from vegetated surfaces proceeds as follows:

- Initial values for canopy air temperature and specific humidity are obtained from

$$T_s = \frac{T_g + \theta_{atm}}{2} \quad (2.245)$$

$$q_s = \frac{q_g + q_{atm}}{2}. \quad (2.246)$$

2. An initial guess for the wind speed V_a is obtained from (2.133) assuming an initial convective velocity $U_c = 0 \text{ m s}^{-1}$ for stable conditions ($\theta_{v, atm} - \theta_{v, s} \geq 0$ as evaluated from (2.159)) and $U_c = 0.5$ for unstable conditions ($\theta_{v, atm} - \theta_{v, s} < 0$).
3. An initial guess for the Monin-Obukhov length L is obtained from the bulk Richardson number using equation and (2.155) and (2.157).
4. Iteration proceeds on the following system of equations:
5. Friction velocity u_* ((2.141), (2.142), (2.143), (2.144))
6. Ratio $\frac{\theta_*}{\theta_{atm} - \theta_s}$ ((2.146), (2.147), (2.148), (2.149))
7. Ratio $\frac{q_*}{q_{atm} - q_s}$ ((2.150), (2.151), (2.152), (2.153))
8. Aerodynamic resistances r_{am} , r_{ah} , and r_{aw} ((2.164), (2.165), (2.166))
9. Magnitude of the wind velocity incident on the leaves U_{av} ((2.226))
10. Leaf boundary layer resistance r_b ((2.245))
11. Aerodynamic resistances r'_{ah} and r'_{aw} ((2.225))
12. Sunlit and shaded stomatal resistances r_s^{sun} and r_s^{sha} (Chapter 2.9)
13. Sensible heat conductances c_a^h , c_g^h , and c_v^h ((2.203), (2.204), (2.205))
14. Latent heat conductances c_a^w , c_v^w , and c_g^w ((2.217), (2.218), (2.219))
15. Sensible heat flux from vegetation H_v ((2.206))
16. Latent heat flux from vegetation λE_v ((2.210))
17. If the latent heat flux has changed sign from the latent heat flux computed at the previous iteration ($\lambda E_v^{n+1} \times \lambda E_v^n < 0$), the latent heat flux is constrained to be 10% of the computed value. The difference between the constrained and computed value ($\Delta_1 = 0.1\lambda E_v^{n+1} - \lambda E_v^n$) is added to the sensible heat flux later.
18. Change in vegetation temperature ΔT_v ((2.238)) and update the vegetation temperature as $T_v^{n+1} = T_v^n + \Delta T_v$. T_v is constrained to change by no more than 1°K in one iteration. If this limit is exceeded, the energy error is

$$\Delta_2 = \vec{S}_v - \vec{L}_v - \frac{\partial \vec{L}_v}{\partial T_v} \Delta T_v - H_v - \frac{\partial H_v}{\partial T_v} \Delta T_v - \lambda E_v - \frac{\partial \lambda E_v}{\partial T_v} \Delta T_v \quad (2.247)$$

where $\Delta T_v = 1$ or -1 . The error Δ_2 is added to the sensible heat flux later.

1. Water vapor flux E_v ((2.242))
2. Transpiration E_v^t ((2.243)) if $\beta_t > 0$, otherwise $E_v^t = 0$
3. The water vapor flux E_v is constrained to be less than or equal to the sum of transpiration E_v^t and the water available from wetted leaves and stems $W_{can}/\Delta t$. The energy error due to this constraint is

$$\Delta_3 = \max \left(0, E_v - E_v^t - \frac{W_{can}}{\Delta t} \right). \quad (2.248)$$

The error $\lambda \Delta_3$ is added to the sensible heat flux later.

1. Sensible heat flux H_v (eq:“). The three energy error terms, Δ_1 , Δ_2 , and $\lambda \Delta_3$ are also added to the sensible heat flux.
2. The saturated vapor pressure e_i (Chapter 2.9), saturated specific humidity $q_{sat}^{T_v}$ and its derivative $\frac{dq_{sat}^{T_v}}{dT_v}$ at the leaf surface (section 2.5.5), are re-evaluated based on the new T_v .
3. Canopy air temperature T_s ((2.202))

4. Canopy air specific humidity q_s ((2.216))
5. Temperature difference $\theta_{atm} - \theta_s$
6. Specific humidity difference $q_{atm} - q_s$
7. Potential temperature scale $\theta_* = \frac{\theta_*}{\theta_{atm} - \theta_s} (\theta_{atm} - \theta_s)$ where $\frac{\theta_*}{\theta_{atm} - \theta_s}$ was calculated earlier in the iteration
8. Humidity scale $q_* = \frac{q_*}{q_{atm} - q_s} (q_{atm} - q_s)$ where $\frac{q_*}{q_{atm} - q_s}$ was calculated earlier in the iteration
9. Virtual potential temperature scale θ_{v*} ((2.126))
10. Wind speed including the convective velocity, V_a ((2.133))
11. Monin-Obukhov length L ((2.158))
12. The iteration is stopped after two or more steps if $\tilde{\Delta}T_v < 0.01$ and $|\lambda E_v^{n+1} - \lambda E_v^n| < 0.1$ where $\tilde{\Delta}T_v = \max(|T_v^{n+1} - T_v^n|, |T_v^n - T_v^{n-1}|)$, or after forty iterations have been carried out.
13. Momentum fluxes τ_x, τ_y ((2.114), (2.115))
14. Sensible heat flux from ground H_g ((2.198))
15. Water vapor flux from ground E_g ((2.211))
16. 2-m height air temperature T_{2m} , specific humidity q_{2m} , relative humidity RH_{2m} ((2.167), (2.168), (2.169))

2.5.4 Update of Ground Sensible and Latent Heat Fluxes

The sensible and water vapor heat fluxes derived above for bare soil and soil beneath canopy are based on the ground surface temperature from the previous time step T_g^n and are used as the surface forcing for the solution of the soil temperature equations (section 2.6.1). This solution yields a new ground surface temperature T_g^{n+1} . The ground sensible and water vapor fluxes are then updated for T_g^{n+1} as

$$H'_g = H_g + (T_g^{n+1} - T_g^n) \frac{\partial H_g}{\partial T_g} \quad (2.249)$$

$$E'_g = E_g + (T_g^{n+1} - T_g^n) \frac{\partial E_g}{\partial T_g} \quad (2.250)$$

where H_g and E_g are the sensible heat and water vapor fluxes derived from equations and for non-vegetated surfaces and equations and for vegetated surfaces using T_g^n . One further adjustment is made to H'_g and E'_g . If the soil moisture in the top snow/soil layer is not sufficient to support the updated ground evaporation, i.e., if $E'_g > 0$ and $f_{evap} < 1$ where

$$f_{evap} = \frac{(w_{ice, snl+1} + w_{liq, snl+1}) / \Delta t}{\sum_{j=1}^{npft} (E'_g)_j (wt)_j} \leq 1, \quad (2.251)$$

an adjustment is made to reduce the ground evaporation accordingly as

$$E''_g = f_{evap} E'_g. \quad (2.252)$$

The term $\sum_{j=1}^{npft} (E'_g)_j (wt)_j$ is the sum of E'_g over all evaporating PFTs where $(E'_g)_j$ is the ground evaporation from the j^{th} PFT on the column, $(wt)_j$ is the relative area of the j^{th} PFT with respect to the column, and $npft$ is the number of PFTs on the column. $w_{ice, snl+1}$ and $w_{liq, snl+1}$ are the ice and liquid water contents (kg m^{-2}) of the top snow/soil layer (Chapter 2.7). Any resulting energy deficit is assigned to sensible heat as

$$H''_g = H_g + \lambda (E'_g - E''_g). \quad (2.253)$$

The ground water vapor flux E_g'' is partitioned into evaporation of liquid water from snow/soil q_{seva} ($\text{kg m}^{-2} \text{ s}^{-1}$), sublimation from snow/soil ice q_{subl} ($\text{kg m}^{-2} \text{ s}^{-1}$), liquid dew on snow/soil q_{sdew} ($\text{kg m}^{-2} \text{ s}^{-1}$), or frost on snow/soil q_{frost} ($\text{kg m}^{-2} \text{ s}^{-1}$) as

$$q_{seva} = \max \left(E_{sno}'' \frac{w_{liq, snl+1}}{w_{ice, snl+1} + w_{liq, snl+1}}, 0 \right) \quad E_{sno}'' \geq 0, w_{ice, snl+1} + w_{liq, snl+1} > 0 \quad (2.254)$$

$$q_{subl} = E_{sno}'' - q_{seva} \quad E_{sno}'' \geq 0 \quad (2.255)$$

$$q_{sdew} = |E_{sno}''| \quad E_{sno}'' < 0 \text{ and } T_g \geq T_f \quad (2.256)$$

$$q_{frost} = |E_{sno}''| \quad E_{sno}'' < 0 \text{ and } T_g < T_f. \quad (2.257)$$

The loss or gain in snow mass due to q_{seva} , q_{subl} , q_{sdew} , and q_{frost} on a snow surface are accounted for during the snow hydrology calculations (Chapter 2.8). The loss of soil and surface water due to q_{seva} is accounted for in the calculation of infiltration (section 2.7.2), while losses or gains due to q_{subl} , q_{sdew} , and q_{frost} on a soil surface are accounted for following the sub-surface drainage calculations (section 2.7.5).

The ground heat flux G is calculated as

$$G = \vec{S}_g - \vec{L}_g - H_g - \lambda E_g \quad (2.258)$$

where \vec{S}_g is the solar radiation absorbed by the ground (section 2.4.1), \vec{L}_g is the net longwave radiation absorbed by the ground (section 2.4.2)

$$\vec{L}_g = L_g \uparrow -\delta_{veg} \varepsilon_g L_v \downarrow - (1 - \delta_{veg}) \varepsilon_g L_{atm} \downarrow + 4 \varepsilon_g \sigma (T_g^n)^3 (T_g^{n+1} - T_g^n), \quad (2.259)$$

where

$$L_g \uparrow = \varepsilon_g \sigma \left[(1 - f_{sno} - f_{h2osfc}) (T_1^n)^4 + f_{sno} (T_{sno}^n)^4 + f_{h2osfc} (T_{h2osfc}^n)^4 \right] \quad (2.260)$$

and H_g and λE_g are the sensible and latent heat fluxes after the adjustments described above.

When converting ground water vapor flux to an energy flux, the term λ is arbitrarily assumed to be

$$\lambda = \begin{cases} \lambda_{sub} & \text{if } w_{liq, snl+1} = 0 \text{ and } w_{ice, snl+1} > 0 \\ \lambda_{vap} & \text{otherwise} \end{cases} \quad (2.261)$$

where λ_{sub} and λ_{vap} are the latent heat of sublimation and vaporization, respectively (J kg^{-1}) (Table 2.7). When converting vegetation water vapor flux to an energy flux, λ_{vap} is used.

The system balances energy as

$$\vec{S}_g + \vec{S}_v + L_{atm} \downarrow - L \uparrow - H_v - H_g - \lambda_{vap} E_v - \lambda E_g - G = 0. \quad (2.262)$$

2.5.5 Saturation Vapor Pressure

Saturation vapor pressure e_{sat}^T (Pa) and its derivative $\frac{de_{sat}^T}{dT}$, as a function of temperature T ($^\circ\text{C}$), are calculated from the eighth-order polynomial fits of *Flatau et al. (1992)*

$$e_{sat}^T = 100 [a_0 + a_1 T + \dots + a_n T^n] \quad (2.263)$$

$$\frac{de_{sat}^T}{dT} = 100 [b_0 + b_1 T + \dots + b_n T^n] \quad (2.264)$$

where the coefficients for ice are valid for $-75^\circ\text{C} \leq T < 0^\circ\text{C}$ and the coefficients for water are valid for $0^\circ\text{C} \leq T \leq 100^\circ\text{C}$ (Table 2.17 and Table 2.18). The saturated water vapor specific humidity q_{sat}^T and its derivative $\frac{dq_{sat}^T}{dT}$ are

$$q_{sat}^T = \frac{0.622e_{sat}^T}{P_{atm} - 0.378e_{sat}^T} \quad (2.265)$$

$$\frac{dq_{sat}^T}{dT} = \frac{0.622P_{atm}}{(P_{atm} - 0.378e_{sat}^T)^2} \frac{de_{sat}^T}{dT}. \quad (2.266)$$

Table 2.17: Coefficients for e_{sat}^T

	water	ice
a_0	6.11213476	6.11123516
a_1	$4.44007856 \times 10^{-1}$	$5.03109514 \times 10^{-1}$
a_2	$1.43064234 \times 10^{-2}$	$1.88369801 \times 10^{-2}$
a_3	$2.64461437 \times 10^{-4}$	$4.20547422 \times 10^{-4}$
a_4	$3.05903558 \times 10^{-6}$	$6.14396778 \times 10^{-6}$
a_5	$1.96237241 \times 10^{-8}$	$6.02780717 \times 10^{-8}$
a_6	$8.92344772 \times 10^{-11}$	$3.87940929 \times 10^{-10}$
a_7	$-3.73208410 \times 10^{-13}$	$1.49436277 \times 10^{-12}$
a_8	$2.09339997 \times 10^{-16}$	$2.62655803 \times 10^{-15}$

Table 2.18: Coefficients for $\frac{de_{sat}^T}{dT}$

	water	ice
b_0	$4.44017302 \times 10^{-1}$	$5.03277922 \times 10^{-1}$
b_1	$2.86064092 \times 10^{-2}$	$3.77289173 \times 10^{-2}$
b_2	$7.94683137 \times 10^{-4}$	$1.26801703 \times 10^{-3}$
b_3	$1.21211669 \times 10^{-5}$	$2.49468427 \times 10^{-5}$
b_4	$1.03354611 \times 10^{-7}$	$3.13703411 \times 10^{-7}$
b_5	$4.04125005 \times 10^{-10}$	$2.57180651 \times 10^{-9}$
b_6	$-7.88037859 \times 10^{-13}$	$1.33268878 \times 10^{-11}$
b_7	$-1.14596802 \times 10^{-14}$	$3.94116744 \times 10^{-14}$
b_8	$3.81294516 \times 10^{-17}$	$4.98070196 \times 10^{-17}$

2.6 Soil and Snow Temperatures

The first law of heat conduction is

$$F = -\lambda \nabla T \quad (2.267)$$

where F is the amount of heat conducted across a unit cross-sectional area in unit time (W m^{-2}), λ is thermal conductivity ($\text{W m}^{-1} \text{ K}^{-1}$), and ∇T is the spatial gradient of temperature (K m^{-1}). In one-dimensional form

$$F_z = -\lambda \frac{\partial T}{\partial z} \quad (2.268)$$

where z is in the vertical direction (m) and is positive downward and F_z is positive upward. To account for non-steady or transient conditions, the principle of energy conservation in the form of the continuity equation is invoked as

$$c \frac{\partial T}{\partial t} = - \frac{\partial F_z}{\partial z} \quad (2.269)$$

where c is the volumetric snow/soil heat capacity ($\text{J m}^{-3} \text{ K}^{-1}$) and t is time (s). Combining equations and yields the second law of heat conduction in one-dimensional form

$$c \frac{\partial T}{\partial t} = \frac{\partial}{\partial z} \left[\lambda \frac{\partial T}{\partial z} \right]. \quad (2.270)$$

This equation is solved numerically to calculate the soil, snow, and surface water temperatures for a fifteen-layer soil column with up to five overlying layers of snow and a single surface water layer with the boundary conditions of h as the heat flux into the top soil, snow, and surface water layers from the overlying atmosphere (section 2.6.1) and zero heat flux at the bottom of the soil column. The temperature profile is calculated first without phase change and then readjusted for phase change (section 2.6.2).

2.6.1 Numerical Solution

The soil column is discretized into 25 layers (section 2.2.2) where $N_{levgrnd} = 25$ is the number of soil layers (Table 2.3).

The overlying snow pack is modeled with up to five layers depending on the total snow depth. The layers from top to bottom are indexed in the Fortran code as $i = -4, -3, -2, -1, 0$, which permits the accumulation or ablation of snow at the top of the snow pack without renumbering the layers. Layer $i = 0$ is the snow layer next to the soil surface and layer $i = snl + 1$ is the top layer, where the variable snl is the negative of the number of snow layers. The number of

snow layers and the thickness of each layer is a function of snow depth z_{sno} (m) as follows.

$$\begin{cases}
 \left. \begin{array}{l} snl = -1 \\ \Delta z_0 = z_{sno} \end{array} \right\} & \text{for } 0.01 \leq z_{sno} \leq 0.03 \\
 \left. \begin{array}{l} snl = -2 \\ \Delta z_{-1} = z_{sno}/2 \\ \Delta z_0 = \Delta z_{-1} \end{array} \right\} & \text{for } 0.03 < z_{sno} \leq 0.04 \\
 \left. \begin{array}{l} snl = -2 \\ \Delta z_{-1} = 0.02 \\ \Delta z_0 = z_{sno} - \Delta z_{-1} \\ snl = -3 \\ \Delta z_{-2} = 0.02 \\ \Delta z_{-1} = (z_{sno} - 0.02)/2 \\ \Delta z_0 = \Delta z_{-1} \end{array} \right\} & \text{for } 0.04 < z_{sno} \leq 0.07 \\
 \left. \begin{array}{l} snl = -3 \\ \Delta z_{-2} = 0.02 \\ \Delta z_{-1} = 0.05 \\ \Delta z_0 = z_{sno} - \Delta z_{-2} - \Delta z_{-1} \end{array} \right\} & \text{for } 0.07 < z_{sno} \leq 0.12 \\
 \left. \begin{array}{l} snl = -4 \\ \Delta z_{-3} = 0.02 \\ \Delta z_{-2} = 0.05 \\ \Delta z_{-1} = (z_{sno} - \Delta z_{-3} - \Delta z_{-2})/2 \\ \Delta z_0 = \Delta z_{-1} \end{array} \right\} & \text{for } 0.12 < z_{sno} \leq 0.18 \\
 \left. \begin{array}{l} snl = -4 \\ \Delta z_{-3} = 0.02 \\ \Delta z_{-2} = 0.05 \\ \Delta z_{-1} = 0.11 \\ \Delta z_0 = z_{sno} - \Delta z_{-3} - \Delta z_{-2} - \Delta z_{-1} \end{array} \right\} & \text{for } 0.18 < z_{sno} \leq 0.29 \\
 \left. \begin{array}{l} snl = -5 \\ \Delta z_{-4} = 0.02 \\ \Delta z_{-3} = 0.05 \\ \Delta z_{-2} = 0.11 \\ \Delta z_{-1} = (z_{sno} - \Delta z_{-4} - \Delta z_{-3} - \Delta z_{-2})/2 \\ \Delta z_0 = \Delta z_{-1} \end{array} \right\} & \text{for } 0.29 < z_{sno} \leq 0.41 \\
 \left. \begin{array}{l} snl = -5 \\ \Delta z_{-4} = 0.02 \\ \Delta z_{-3} = 0.05 \\ \Delta z_{-2} = 0.11 \\ \Delta z_{-1} = 0.23 \\ \Delta z_0 = z_{sno} - \Delta z_{-4} - \Delta z_{-3} - \Delta z_{-2} - \Delta z_{-1} \end{array} \right\} & \text{for } 0.41 < z_{sno} \leq 0.64 \\
 \left. \begin{array}{l} snl = -5 \\ \Delta z_{-4} = 0.02 \\ \Delta z_{-3} = 0.05 \\ \Delta z_{-2} = 0.11 \\ \Delta z_{-1} = 0.23 \\ \Delta z_0 = z_{sno} - \Delta z_{-4} - \Delta z_{-3} - \Delta z_{-2} - \Delta z_{-1} \end{array} \right\} & \text{for } 0.64 < z_{sno}
 \end{cases}$$

The node depths, which are located at the midpoint of the snow layers, and the layer interfaces are both referenced from the soil surface and are defined as negative values

$$z_i = z_{h,i} - 0.5\Delta z_i \quad i = snl + 1, \dots, 0 \quad (2.271)$$

$$z_{h,i} = z_{h,i+1} - \Delta z_{i+1} \quad i = snl, \dots, -1. \quad (2.272)$$

Note that $z_{h,0}$, the interface between the bottom snow layer and the top soil layer, is zero. Thermal properties (i.e., temperature T_i [K]; thermal conductivity λ_i [$\text{W m}^{-1} \text{K}^{-1}$]; volumetric heat capacity c_i [$\text{J m}^{-3} \text{K}^{-1}$]) are defined for soil layers at the node depths (Figure 2.6) and for snow layers at the layer midpoints. When present, snow occupies a fraction of a grid cell's area, therefore snow depth represents the thickness of the snowpack averaged over only the snow covered area. The grid cell average snow depth is related to the depth of the snow covered area as $\bar{z}_{sno} = f_{sno} z_{sno}$. By default, the grid cell average snow depth is written to the history file.

The heat flux F_i (W m^{-2}) from layer i to layer $i + 1$ is

$$F_i = -\lambda[z_{h,i}] \left(\frac{T_i - T_{i+1}}{z_{i+1} - z_i} \right) \quad (2.273)$$

where the thermal conductivity at the interface $\lambda[z_{h,i}]$ is

$$\lambda[z_{h,i}] = \begin{cases} \frac{\lambda_i \lambda_{i+1} (z_{i+1} - z_i)}{\lambda_i (z_{i+1} - z_{h,i}) + \lambda_{i+1} (z_{h,i} - z_i)} & i = snl + 1, \dots, N_{levgrnd} - 1 \\ 0 & i = N_{levgrnd} \end{cases}. \quad (2.274)$$

These equations are derived, with reference to Figure 2.6, assuming that the heat flux from i (depth z_i) to the interface between i and $i + 1$ (depth $z_{h,i}$) equals the heat flux from the interface to $i + 1$ (depth z_{i+1}), i.e.,

$$-\lambda_i \frac{T_i - T_m}{z_{h,i} - z_i} = -\lambda_{i+1} \frac{T_m - T_{i+1}}{z_{i+1} - z_{h,i}} \quad (2.275)$$

where T_m is the temperature at the interface of layers i and $i + 1$.

Shown are three soil layers, $i - 1$, i , and $i + 1$. The thermal conductivity λ , specific heat capacity c , and temperature T are defined at the layer node depth z . T_m is the interface temperature. The thermal conductivity $\lambda[z_h]$ is defined at the interface of two layers

z_h . The layer thickness is Δz . The heat fluxes F_{i-1} and F_i are defined as positive upwards.

The energy balance for the i^{th} layer is

$$\frac{c_i \Delta z_i}{\Delta t} (T_i^{n+1} - T_i^n) = -F_{i-1} + F_i \quad (2.276)$$

where the superscripts n and $n + 1$ indicate values at the beginning and end of the time step, respectively, and Δt is the time step (s). This equation is solved using the Crank–Nicholson method, which combines the explicit method with fluxes evaluated at n (F_{i-1}^n, F_i^n) and the implicit method with fluxes evaluated at $n + 1$ (F_{i-1}^{n+1}, F_i^{n+1})

$$\frac{c_i \Delta z_i}{\Delta t} (T_i^{n+1} - T_i^n) = \alpha (-F_{i-1}^n + F_i^n) + (1 - \alpha) (-F_{i-1}^{n+1} + F_i^{n+1}) \quad (2.277)$$

where $\alpha = 0.5$, resulting in a tridiagonal system of equations

$$r_i = a_i T_{i-1}^{n+1} + b_i T_i^{n+1} + c_i T_{i+1}^{n+1} \quad (2.278)$$

where a_i , b_i , and c_i are the subdiagonal, diagonal, and superdiagonal elements in the tridiagonal matrix and r_i is a column vector of constants. When surface water is present, the equation for the top soil layer has an additional term representing the surface water temperature; this results in a four element band-diagonal system of equations.

For the top soil layer $i = 1$, top snow layer $i = snl + 1$, or surface water layer, the heat flux from the overlying atmosphere h (W m^{-2} , defined as positive into the surface) is

$$h^{n+1} = -\alpha F_{i-1}^n - (1 - \alpha) F_{i-1}^{n+1}. \quad (2.279)$$

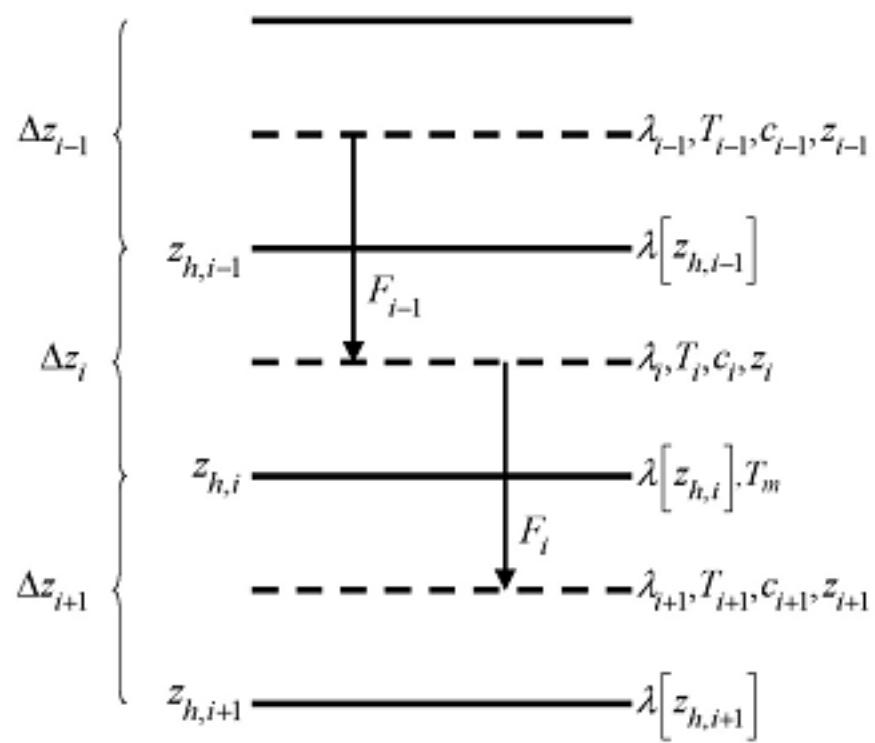


Figure 2.6: Schematic diagram of numerical scheme used to solve for soil temperature.

The energy balance for these layers is then

$$\frac{c_i \Delta z_i}{\Delta t} (T_i^{n+1} - T_i^n) = h^{n+1} + \alpha F_i^n + (1 - \alpha) F_i^{n+1}. \quad (2.280)$$

The heat flux h at $n + 1$ may be approximated as follows

$$h^{n+1} = h^n + \frac{\partial h}{\partial T_i} (T_i^{n+1} - T_i^n). \quad (2.281)$$

The resulting equations are then

$$\begin{aligned} \frac{c_i \Delta z_i}{\Delta t} (T_i^{n+1} - T_i^n) &= h^n + \frac{\partial h}{\partial T_i} (T_i^{n+1} - T_i^n) \\ &\quad - \alpha \frac{\lambda[z_{h,i}] (T_i^n - T_{i+1}^n)}{z_{i+1} - z_i} - (1 - \alpha) \frac{\lambda[z_{h,i}] (T_i^{n+1} - T_{i+1}^{n+1})}{z_{i+1} - z_i} \end{aligned} \quad (2.282)$$

For the top snow layer, $i = snl + 1$, the coefficients are

$$a_i = 0 \quad (2.283)$$

$$b_i = 1 + \frac{\Delta t}{c_i \Delta z_i} \left[(1 - \alpha) \frac{\lambda[z_{h,i}]}{z_{i+1} - z_i} - \frac{\partial h}{\partial T_i} \right] \quad (2.284)$$

$$c_i = -(1 - \alpha) \frac{\Delta t}{c_i \Delta z_i} \frac{\lambda[z_{h,i}]}{z_{i+1} - z_i} \quad (2.285)$$

$$r_i = T_i^n + \frac{\Delta t}{c_i \Delta z_i} \left[h_{sno}^n - \frac{\partial h}{\partial T_i} T_i^n + \alpha F_i \right] \quad (2.286)$$

where

$$F_i = -\lambda[z_{h,i}] \left(\frac{T_i^n - T_{i+1}^n}{z_{i+1} - z_i} \right). \quad (2.287)$$

The heat flux into the snow surface from the overlying atmosphere h is

$$h = \vec{S}_{sno} - \vec{L}_{sno} - H_{sno} - \lambda E_{sno} \quad (2.288)$$

where \vec{S}_{sno} is the solar radiation absorbed by the top snow layer (section 2.3.2), \vec{L}_{sno} is the longwave radiation absorbed by the snow (positive toward the atmosphere) (section 2.4.2), H_{sno} is the sensible heat flux from the snow (Chapter 2.5), and λE_{sno} is the latent heat flux from the snow (Chapter 2.5). The partial derivative of the heat flux h with respect to temperature is

$$\frac{\partial h}{\partial T} = -\frac{\partial \vec{L}}{\partial T} - \frac{\partial H}{\partial T} - \frac{\partial \lambda E}{\partial T} \quad (2.289)$$

where the partial derivative of the net longwave radiation is

$$\frac{\partial \vec{L}}{\partial T} = 4\varepsilon_g \sigma (T^n)^3 \quad (2.290)$$

and the partial derivatives of the sensible and latent heat fluxes are given by equations and for non-vegetated surfaces, and by equations and for vegetated surfaces. σ is the Stefan-Boltzmann constant ($\text{W m}^{-2} \text{K}^{-4}$) (Table 2.7) and ε_g is the ground emissivity (section 2.4.2). For purposes of computing h and $\frac{\partial h}{\partial T_g}$, the term λ is arbitrarily assumed to be

$$\lambda = \begin{cases} \lambda_{sub} & \text{if } w_{liq,snl+1} = 0 \text{ and } w_{ice,snl+1} > 0 \\ \lambda_{vap} & \text{otherwise} \end{cases} \quad (2.291)$$

where λ_{sub} and λ_{vap} are the latent heat of sublimation and vaporization, respectively (J kg^{-1}) (Table 2.7), and $w_{liq, snl+1}$ and $w_{ice, snl+1}$ are the liquid water and ice contents of the top snow/soil layer, respectively (kg m^{-2}) (Chapter 2.7).

For the top soil layer, $i = 1$, the coefficients are

$$a_i = -f_{sno} (1 - \alpha) \frac{\Delta t}{c_i \Delta z_i} \frac{\lambda [z_{h, i-1}]}{z_i - z_{i-1}} \quad (2.292)$$

$$b_i = 1 + (1 - \alpha) \frac{\Delta t}{c_i \Delta z_i} \left[f_{sno} \frac{\lambda [z_{h, i-1}]}{z_i - z_{i-1}} + \frac{\lambda [z_{h, i}]}{z_{i+1} - z_i} \right] - (1 - f_{sno}) \frac{\Delta t}{c_i \Delta z_i} \frac{\partial h}{\partial T} \quad (2.293)$$

$$c_i = -(1 - \alpha) \frac{\Delta t}{c_i \Delta z_i} \frac{\lambda [z_{h, i}]}{z_{i+1} - z_i} \quad (2.294)$$

$$r_i = T_i^n + \frac{\Delta t}{c_i \Delta z_i} \left[(1 - f_{sno}) \left(h_{soil}^n - \frac{\partial h}{\partial T} T_i^n \right) + \alpha (F_i - f_{sno} F_{i-1}) \right] \quad (2.295)$$

The heat flux into the soil surface from the overlying atmosphere h is

$$h = \vec{S}_{soil} - \vec{L}_{soil} - H_{soil} - \lambda E_{soil} \quad (2.296)$$

It can be seen that when no snow is present ($f_{sno} = 0$), the expressions for the coefficients of the top soil layer have the same form as those for the top snow layer.

The surface snow/soil layer temperature computed in this way is the layer-averaged temperature and hence has somewhat reduced diurnal amplitude compared with surface temperature. An accurate surface temperature is provided that compensates for this effect and numerical error by tuning the heat capacity of the top layer (through adjustment of the layer thickness) to give an exact match to the analytic solution for diurnal heating. The top layer thickness for $i = snl + 1$ is given by

$$\Delta z_{i*} = 0.5 [z_i - z_{h, i-1} + c_a (z_{i+1} - z_{h, i-1})] \quad (2.297)$$

where c_a is a tunable parameter, varying from 0 to 1, and is taken as 0.34 by comparing the numerical solution with the analytic solution (Z.-L. Yang 1998, unpublished manuscript). Δz_{i*} is used in place of Δz_i for $i = snl + 1$ in equations -. The top snow/soil layer temperature computed in this way is the ground surface temperature T_g^{n+1} .

The boundary condition at the bottom of the snow/soil column is zero heat flux, $F_i = 0$, resulting in, for $i = N_{levgrnd}$,

$$\frac{c_i \Delta z_i}{\Delta t} (T_i^{n+1} - T_i^n) = \alpha \frac{\lambda [z_{h, i-1}] (T_{i-1}^n - T_i^n)}{z_i - z_{i-1}} + (1 - \alpha) \frac{\lambda [z_{h, i-1}] (T_{i-1}^{n+1} - T_i^{n+1})}{z_i - z_{i-1}} \quad (2.298)$$

$$a_i = -(1 - \alpha) \frac{\Delta t}{c_i \Delta z_i} \frac{\lambda [z_{h, i-1}]}{z_i - z_{i-1}} \quad (2.299)$$

$$b_i = 1 + (1 - \alpha) \frac{\Delta t}{c_i \Delta z_i} \frac{\lambda [z_{h, i-1}]}{z_i - z_{i-1}} \quad (2.300)$$

$$c_i = 0 \quad (2.301)$$

$$r_i = T_i^n - \alpha \frac{\Delta t}{c_i \Delta z_i} F_{i-1} \quad (2.302)$$

where

$$F_{i-1} = -\frac{\lambda[z_h, i-1]}{z_i - z_{i-1}} (T_{i-1}^n - T_i^n). \quad (2.303)$$

For the interior snow/soil layers, $snl + 1 < i < N_{levgrnd}$, excluding the top soil layer,

$$\frac{c_i \Delta z_i}{\Delta t} (T_i^{n+1} - T_i^n) = -\alpha \frac{\lambda[z_h, i] (T_i^n - T_{i+1}^n)}{z_{i+1} - z_i} + \alpha \frac{\lambda[z_h, i-1] (T_{i-1}^n - T_i^n)}{z_i - z_{i-1}} \quad (2.304)$$

$$a_i = -(1 - \alpha) \frac{\Delta t}{c_i \Delta z_i} \frac{\lambda[z_h, i-1]}{z_i - z_{i-1}} \quad (2.305)$$

$$b_i = 1 + (1 - \alpha) \frac{\Delta t}{c_i \Delta z_i} \left[\frac{\lambda[z_h, i-1]}{z_i - z_{i-1}} + \frac{\lambda[z_h, i]}{z_{i+1} - z_i} \right] \quad (2.306)$$

$$c_i = -(1 - \alpha) \frac{\Delta t}{c_i \Delta z_i} \frac{\lambda[z_h, i]}{z_{i+1} - z_i} \quad (2.307)$$

$$r_i = T_i^n + \alpha \frac{\Delta t}{c_i \Delta z_i} (F_i - F_{i-1}) + \frac{\Delta t}{c_i \Delta z_i} \vec{S}_{g,i}. \quad (2.308)$$

where $\vec{S}_{g,i}$ is the absorbed solar flux in layer i (section 2.3.2).

When surface water exists, the following top soil layer coefficients are modified

$$b_i = 1 + (1 - \alpha) \frac{\Delta t}{c_i \Delta z_i} \left[f_{h2osfc} \frac{\lambda_{h2osfc}}{z_i - z_{h2osfc}} + f_{sno} \frac{\lambda[z_h, i-1]}{z_i - z_{i-1}} + \frac{\lambda[z_h, i]}{z_{i+1} - z_i} \right] - (1 - f_{sno} - f_{h2osfc}) \frac{\Delta t}{c_i \Delta z_i} \frac{\partial h}{\partial T} \quad (2.309)$$

$$r_i = T_i^n + \frac{\Delta t}{c_i \Delta z_i} \left[(1 - f_{sno} - f_{h2osfc}) \left(h_{soil}^n - \frac{\partial h}{\partial T} T_i^n \right) + \alpha \left(F_i - f_{sno} F_{i-1} + f_{h2osfc} \frac{\lambda_{h2osfc}}{z_1 - z_{h2osfc}} (T_1 - T_{h2osfc}) \right) \right] \quad (2.310)$$

$$d_i = -f_{h2osfc} (1 - \alpha) \frac{\Delta t}{c_i \Delta z_i} \left[\frac{\lambda_{h2osfc}}{z_i - z_{h2osfc}} \right] \quad (2.311)$$

where d_i is an additional coefficient representing the heat flux from the surface water layer. The surface water layer coefficients are

$$a_{h2osfc} = 0 \quad (2.312)$$

$$b_{h2osfc} = 1 + \frac{\Delta t}{c_{h2osfc} \Delta z_{h2osfc}} \left[(1 - \alpha) \frac{\lambda_{h2osfc}}{z_1 - z_{h2osfc}} - \frac{\partial h}{\partial T} \right] \quad (2.313)$$

$$c_{h2osfc} = -(1 - \alpha) \frac{\Delta t}{c_{h2osfc} \Delta z_{h2osfc}} \frac{\lambda_{h2osfc}}{z_1 - z_{h2osfc}} \quad (2.314)$$

$$r_{h2osfc} = T_{h2osfc}^n + \frac{\Delta t}{c_i \Delta z_i} \left[h_{h2osfc}^n - \frac{\partial h}{\partial T} T_{h2osfc}^n + \alpha \frac{\lambda_{h2osfc}}{z_1 - z_{h2osfc}} (T_1 - T_{h2osfc}) \right] \quad (2.315)$$

2.6.2 Phase Change

Soil and Snow Layers

Upon update, the snow/soil temperatures are evaluated to determine if phase change will take place as

$$T_i^{n+1} > T_f \text{ and } w_{ice, i} > 0 \quad i = snl + 1, \dots, N_{levgrnd} \quad \text{melting} \quad (2.316)$$

$$\begin{aligned} T_i^{n+1} &< T_f \text{ and } w_{liq, i} > 0 & i = snl + 1, \dots, 0 \\ T_i^{n+1} &< T_f \text{ and } w_{liq, i} > w_{liq, max, i} & i = 1, \dots, N_{levgrnd} \end{aligned} \quad \text{freezing} \quad (2.317)$$

where T_i^{n+1} is the soil layer temperature after solution of the tridiagonal equation set, $w_{ice, i}$ and $w_{liq, i}$ are the mass of ice and liquid water (kg m^{-2}) in each snow/soil layer, respectively, and T_f is the freezing temperature of water (K) (Table 2.7). For the freezing process in soil layers, the concept of supercooled soil water from [Niu and Yang \(2006\)](#) is adopted. The supercooled soil water is the liquid water that coexists with ice over a wide range of temperatures below freezing and is implemented through a freezing point depression equation

$$w_{liq, max, i} = \Delta z_i \theta_{sat, i} \left[\frac{10^3 L_f (T_f - T_i)}{g T_i \psi_{sat, i}} \right]^{-1/B_i} \quad T_i < T_f \quad (2.318)$$

where $w_{liq, max, i}$ is the maximum liquid water in layer i (kg m^{-2}) when the soil temperature T_i is below the freezing temperature T_f , L_f is the latent heat of fusion (J kg^{-1}) (Table 2.7), g is the gravitational acceleration (m s^{-2}) (Table 2.7), and $\psi_{sat, i}$ and B_i are the soil texture-dependent saturated matric potential (mm) and [Clapp and Hornberger \(1978\)](#) exponent (section 2.7.3).

For the special case when snow is present (snow mass $W_{sno} > 0$) but there are no explicit snow layers ($snl = 0$) (i.e., there is not enough snow present to meet the minimum snow depth requirement of 0.01 m), snow melt will take place for soil layer $i = 1$ if the soil layer temperature is greater than the freezing temperature ($T_1^{n+1} > T_f$).

The rate of phase change is assessed from the energy excess (or deficit) needed to change T_i to freezing temperature, T_f . The excess or deficit of energy H_i (W m^{-2}) is determined as follows

$$H_i = \begin{cases} \frac{\partial h}{\partial T} (T_f - T_i^n) - \frac{c_i \Delta z_i}{\Delta t} (T_f - T_i^n) & i = snl + 1 \\ (1 - f_{sno} - f_{h2osfc}) \frac{\partial h}{\partial T} (T_f - T_i^n) - \frac{c_i \Delta z_i}{\Delta t} (T_f - T_i^n) & i = 1 \\ -\frac{c_i \Delta z_i}{\Delta t} (T_f - T_i^n) & i \neq \{1, snl + 1\} \end{cases}. \quad (2.319)$$

If the melting criteria is met (2.316) and $H_m = \frac{H_i \Delta t}{L_f} > 0$, then the ice mass is readjusted as

$$w_{ice, i}^{n+1} = w_{ice, i}^n - H_m \geq 0 \quad i = snl + 1, \dots, N_{levgrnd}. \quad (2.320)$$

If the freezing criteria is met (2.317) and $H_m < 0$, then the ice mass is readjusted for $i = snl + 1, \dots, 0$ as

$$w_{ice, i}^{n+1} = \min(w_{liq, i}^n + w_{ice, i}^n, w_{ice, i}^n - H_m) \quad (2.321)$$

and for $i = 1, \dots, N_{levgrnd}$ as

$$w_{ice, i}^{n+1} = \begin{cases} \min(w_{liq, i}^n + w_{ice, i}^n - w_{liq, max, i}^n, w_{ice, i}^n - H_m) & w_{liq, i}^n + w_{ice, i}^n \geq w_{liq, max, i}^n \\ 0 & w_{liq, i}^n + w_{ice, i}^n < w_{liq, max, i}^n \end{cases}. \quad (2.322)$$

Liquid water mass is readjusted as

$$w_{liq, i}^{n+1} = w_{liq, i}^n + w_{ice, i}^n - w_{ice, i}^{n+1} \geq 0. \quad (2.323)$$

Because part of the energy H_i may not be consumed in melting or released in freezing, the energy is recalculated as

$$H_{i*} = H_i - \frac{L_f (w_{ice, i}^n - w_{ice, i}^{n+1})}{\Delta t} \quad (2.324)$$

and this energy is used to cool or warm the snow/soil layer (if $|H_{i*}| > 0$) as

$$T_i^{n+1} = \begin{cases} T_f + \frac{\Delta t}{c_i \Delta z_i} H_{i*} / \left(1 - \frac{\Delta t}{c_i \Delta z_i} \frac{\partial h}{\partial T} \right) & i = snl + 1 \\ T_f + \frac{\Delta t}{c_i \Delta z_i} H_{i*} / \left(1 - (1 - f_{sno} - f_{h2osfc}) \frac{\Delta t}{c_i \Delta z_i} \frac{\partial h}{\partial T} \right) & i = 1 \\ T_f + \frac{\Delta t}{c_i \Delta z_i} H_{i*} & i \neq \{1, snl + 1\} \end{cases}. \quad (2.325)$$

For the special case when snow is present ($W_{sno} > 0$), there are no explicit snow layers ($snl = 0$), and $\frac{H_1 \Delta t}{L_f} > 0$ (melting), the snow mass W_{sno} (kg m^{-2}) is reduced according to

$$W_{sno}^{n+1} = W_{sno}^n - \frac{H_1 \Delta t}{L_f} \geq 0. \quad (2.326)$$

The snow depth is reduced proportionally

$$z_{sno}^{n+1} = \frac{W_{sno}^{n+1}}{W_{sno}^n} z_{sno}^n. \quad (2.327)$$

Again, because part of the energy may not be consumed in melting, the energy for the surface soil layer $i = 1$ is recalculated as

$$H_{1*} = H_1 - \frac{L_f (W_{sno}^n - W_{sno}^{n+1})}{\Delta t}. \quad (2.328)$$

If there is excess energy ($H_{1*} > 0$), this energy becomes available to the top soil layer as

$$H_1 = H_{1*}. \quad (2.329)$$

The ice mass, liquid water content, and temperature of the top soil layer are then determined from (2.320), (2.323), and (2.325) using the recalculated energy from (2.329). Snow melt M_{1S} ($\text{kg m}^{-2} \text{s}^{-1}$) and phase change energy $E_{p,1S}$ (W m^{-2}) for this special case are

$$M_{1S} = \frac{W_{sno}^n - W_{sno}^{n+1}}{\Delta t} \geq 0 \quad (2.330)$$

$$E_{p,1S} = L_f M_{1S}. \quad (2.331)$$

The total energy of phase change E_p (W m^{-2}) for the snow/soil column is

$$E_p = E_{p,1S} + \sum_{i=snl+1}^{N_{levgrnd}} E_{p,i} \quad (2.332)$$

where

$$E_{p,i} = L_f \frac{(w_{ice,i}^n - w_{ice,i}^{n+1})}{\Delta t}. \quad (2.333)$$

The total snow melt M ($\text{kg m}^{-2} \text{s}^{-1}$) is

$$M = M_{1S} + \sum_{i=snl+1}^{i=0} M_i \quad (2.334)$$

where

$$M_i = \frac{(w_{ice,i}^n - w_{ice,i}^{n+1})}{\Delta t} \geq 0. \quad (2.335)$$

The solution for snow/soil temperatures conserves energy as

$$G - E_p - \sum_{i=snl+1}^{i=N_{levgrnd}} \frac{c_i \Delta z_i}{\Delta t} (T_i^{n+1} - T_i^n) = 0 \quad (2.336)$$

where G is the ground heat flux (section 2.5.4).

Surface Water

Phase change of surface water takes place when the surface water temperature, T_{h2osfc} , becomes less than T_f . The energy available for freezing is

$$H_{h2osfc} = \frac{\partial h}{\partial T} (T_f - T_{h2osfc}^n) - \frac{c_{h2osfc} \Delta z_{h2osfc}}{\Delta t} (T_f - T_{h2osfc}^n) \quad (2.337)$$

where c_{h2osfc} is the volumetric heat capacity of water, and Δz_{h2osfc} is the depth of the surface water layer. If $H_m = \frac{H_{h2osfc} \Delta t}{L_f} > 0$ then H_m is removed from surface water and added to the snow column as ice

$$H_{h2osfc}^{n+1} = H_{h2osfc}^n - H_m \quad (2.338)$$

$$w_{ice,0}^{n+1} = w_{ice,0}^n + H_m \quad (2.339)$$

The snow depth is adjusted to account for the additional ice mass

$$\Delta z_{sno} = \frac{H_m}{\rho_{ice}} \quad (2.340)$$

If H_m is greater than W_{sfc} , the excess heat $\frac{L_f(H_m - W_{sfc})}{\Delta t}$ is used to cool the snow layer.

2.6.3 Soil and Snow Thermal Properties

The thermal properties of the soil are assumed to be a weighted combination of the mineral and organic properties of the soil ([Lawrence and Slater 2008](#)). The soil layer organic matter fraction $f_{om,i}$ is

$$f_{om,i} = \rho_{om,i}/\rho_{om,max}. \quad (2.341)$$

Soil thermal conductivity λ_i ($\text{W m}^{-1} \text{K}^{-1}$) is from [Farouki \(1981\)](#)

$$\lambda_i = \begin{cases} K_{e,i} \lambda_{sat,i} + (1 - K_{e,i}) \lambda_{dry,i} & S_{r,i} > 1 \times 10^{-7} \\ \lambda_{dry,i} & S_{r,i} \leq 1 \times 10^{-7} \end{cases} \quad i = 1, \dots, N_{levsoi} \quad (2.342)$$

$$\lambda_i = \lambda_{bedrock} \quad i = N_{levsoi} + 1, \dots, N_{levrnd}$$

where $\lambda_{sat,i}$ is the saturated thermal conductivity, $\lambda_{dry,i}$ is the dry thermal conductivity, $K_{e,i}$ is the Kersten number, $S_{r,i}$ is the wetness of the soil with respect to saturation, and $\lambda_{bedrock} = 3 \text{ W m}^{-1} \text{K}^{-1}$ is the thermal conductivity assumed for the deep ground layers (typical of saturated granitic rock; [Clauer and Huenges 1995](#)). For glaciers and wetlands,

$$\lambda_i = \begin{cases} \lambda_{liq,i} & T_i \geq T_f \\ \lambda_{ice,i} & T_i < T_f \end{cases} \quad (2.343)$$

where λ_{liq} and λ_{ice} are the thermal conductivities of liquid water and ice, respectively ([Table 2.7](#)). The saturated thermal conductivity $\lambda_{sat,i}$ ($\text{W m}^{-1} \text{K}^{-1}$) depends on the thermal conductivities of the soil solid, liquid water, and ice constituents

$$\lambda_{sat} = \lambda_s^{1-\theta_{sat}} \lambda_{liq}^{\frac{\theta_{liq}}{\theta_{liq}+\theta_{ice}}} \theta_{sat} \lambda_{ice}^{\theta_{sat} \left(1 - \frac{\theta_{liq}}{\theta_{liq}+\theta_{ice}}\right)} \quad (2.344)$$

where the thermal conductivity of soil solids $\lambda_{s,i}$ varies with the sand, clay, and organic matter content

$$\lambda_{s,i} = (1 - f_{om,i}) \lambda_{s,min,i} + f_{om,i} \lambda_{s,om} \quad (2.345)$$

where the mineral soil solid thermal conductivity $\lambda_{s,\min,i}$ is

$$\lambda_{s,\min,i} = \frac{8.80 (\%sand)_i + 2.92 (\%clay)_i}{(\%sand)_i + (\%clay)_i}, \quad (2.346)$$

and $\lambda_{s,om} = 0.25 \text{ W m}^{-1} \text{ K}^{-1}$ (*Farouki 1981*). $\theta_{sat,i}$ is the volumetric water content at saturation (porosity) (section 2.7.3).

The thermal conductivity of dry soil is

$$\lambda_{dry,i} = (1 - f_{om,i})\lambda_{dry,\min,i} + f_{om,i}\lambda_{dry,om} \quad (2.347)$$

where the thermal conductivity of dry mineral soil $\lambda_{dry,\min,i}$ ($\text{W m}^{-1} \text{ K}^{-1}$) depends on the bulk density $\rho_{d,i} = 2700 (1 - \theta_{sat,i})$ (kg m^{-3}) as

$$\lambda_{dry,\min,i} = \frac{0.135\rho_{d,i} + 64.7}{2700 - 0.947\rho_{d,i}} \quad (2.348)$$

and $\lambda_{dry,om} = 0.05 \text{ W m}^{-1} \text{ K}^{-1}$ (*Farouki 1981*) is the dry thermal conductivity of organic matter. The Kersten number $K_{e,i}$ is a function of the degree of saturation S_r and phase of water

$$K_{e,i} = \begin{cases} \log(S_{r,i}) + 1 \geq 0 & T_i \geq T_f \\ S_{r,i} & T_i < T_f \end{cases} \quad (2.349)$$

where

$$S_{r,i} = \left(\frac{w_{liq,i}}{\rho_{liq}\Delta z_i} + \frac{w_{ice,i}}{\rho_{ice}\Delta z_i} \right) \frac{1}{\theta_{sat,i}} = \frac{\theta_{liq,i} + \theta_{ice,i}}{\theta_{sat,i}} \leq 1. \quad (2.350)$$

Thermal conductivity λ_i ($\text{W m}^{-1} \text{ K}^{-1}$) for snow is from *Jordan (1991)*

$$\lambda_i = \lambda_{air} + (7.75 \times 10^{-5} \rho_{sno,i} + 1.105 \times 10^{-6} \rho_{sno,i}^2) (\lambda_{ice} - \lambda_{air}) \quad (2.351)$$

where λ_{air} is the thermal conductivity of air (Table 2.7) and $\rho_{sno,i}$ is the bulk density of snow (kg m^{-3})

$$\rho_{sno,i} = \frac{w_{ice,i} + w_{liq,i}}{\Delta z_i}. \quad (2.352)$$

The volumetric heat capacity c_i ($\text{J m}^{-3} \text{ K}^{-1}$) for soil is from *de Vries (1963)* and depends on the heat capacities of the soil solid, liquid water, and ice constituents

$$c_i = c_{s,i} (1 - \theta_{sat,i}) + \frac{w_{ice,i}}{\Delta z_i} C_{ice} + \frac{w_{liq,i}}{\Delta z_i} C_{liq} \quad (2.353)$$

where C_{liq} and C_{ice} are the specific heat capacities ($\text{J kg}^{-1} \text{ K}^{-1}$) of liquid water and ice, respectively (Table 2.7). The heat capacity of soil solids $c_{s,i}$ ($\text{J m}^{-3} \text{ K}^{-1}$) is

$$c_{s,i} = (1 - f_{om,i})c_{s,\min,i} + f_{om,i}c_{s,om} \quad (2.354)$$

where the heat capacity of mineral soil solids $c_{s,\min,i}$ ($\text{J m}^{-3} \text{ K}^{-1}$) is

$$c_{s,\min,i} = \begin{cases} \left(\frac{2.128 (\%sand)_i + 2.385 (\%clay)_i}{(\%sand)_i + (\%clay)_i} \right) \times 10^6 & i = 1, \dots, N_{levsoi} \\ c_{s,bedrock} & i = N_{levsoi} + 1, \dots, N_{levgrnd} \end{cases} \quad (2.355)$$

where $c_{s,bedrock} = 2 \times 10^6 \text{ J m}^{-3} \text{ K}^{-1}$ is the heat capacity of bedrock and $c_{s,om} = 2.5 \times 10^6 \text{ J m}^{-3} \text{ K}^{-1}$ (*Farouki 1981*) is the heat capacity of organic matter. For glaciers, wetlands, and snow

$$c_i = \frac{w_{ice,i}}{\Delta z_i} C_{ice} + \frac{w_{liq,i}}{\Delta z_i} C_{liq}. \quad (2.356)$$

For the special case when snow is present ($W_{sno} > 0$) but there are no explicit snow layers ($snl = 0$), the heat capacity of the top layer is a blend of ice and soil heat capacity

$$c_1 = c_1^* + \frac{C_{ice}W_{sno}}{\Delta z_1} \quad (2.357)$$

where c_1^* is calculated from (2.353) or (2.356).

2.7 Hydrology

The model parameterizes interception, throughfall, canopy drip, snow accumulation and melt, water transfer between snow layers, infiltration, evaporation, surface runoff, sub-surface drainage, redistribution within the soil column, and groundwater discharge and recharge to simulate changes in canopy water $\Delta W_{can, liq}$, canopy snow water $\Delta W_{can, sno}$ surface water ΔW_{sfc} , snow water ΔW_{sno} , soil water $\Delta w_{liq, i}$, and soil ice $\Delta w_{ice, i}$, and water in the unconfined aquifer ΔW_a (all in kg m^{-2} or mm of H_2O) (Figure 2.7).

The total water balance of the system is

$$\Delta W_{can, liq} + \Delta W_{can, sno} + \Delta W_{sfc} + \Delta W_{sno} + \sum_{i=1}^{N_{levsoi}} (\Delta w_{liq, i} + \Delta w_{ice, i}) + \Delta W_a = \left(q_{rain} + q_{sno} - E_v - E_g - q_{over} - q_{h2osfc} - q_{drai} - q_{rgwl} - q_{snwcp, ice} \right) \Delta t \quad (2.358)$$

where q_{rain} is the liquid part of precipitation, q_{sno} is the solid part of precipitation, E_v is ET from vegetation (Chapter 2.5), E_g is ground evaporation (Chapter 2.5), q_{over} is surface runoff (section 2.7.2), q_{h2osfc} is runoff from surface water storage (section 2.7.2), q_{drai} is sub-surface drainage (section 2.7.5), q_{rgwl} and $q_{snwcp, ice}$ are liquid and solid runoff from glaciers, wetlands, and lakes, and runoff from other surface types due to snow capping (section 2.7.6) (all in $\text{kg m}^{-2} \text{s}^{-1}$), N_{levsoi} is the number of soil layers (note that hydrology calculations are only done over soil layers 1 to N_{levsoi} ; ground levels $N_{levsoi} + 1$ to $N_{levgrnd}$ are currently hydrologically inactive; (Lawrence et al. 2008) and Δt is the time step (s).

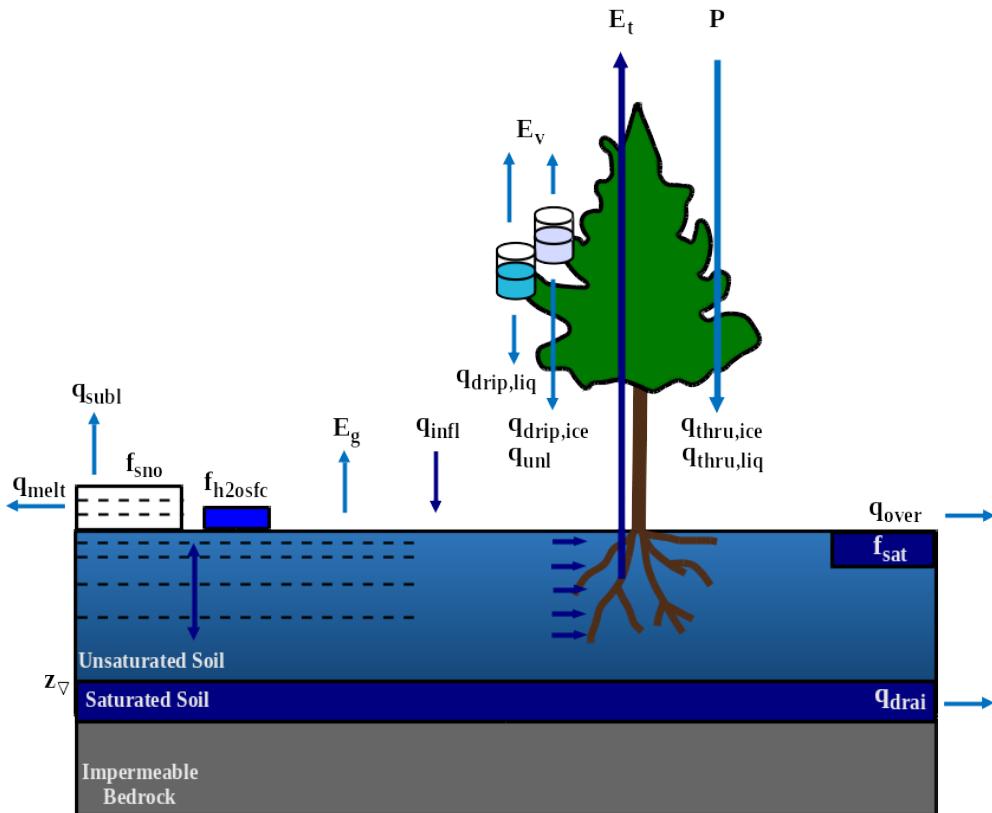


Figure 2.7: Hydrologic processes represented in CLM.

2.7.1 Canopy Water

Liquid precipitation is either intercepted by the canopy, falls directly to the snow/soil surface (throughfall), or drips off the vegetation (canopy drip). Solid precipitation is treated similarly, with the addition of unloading of previously intercepted snow. Interception by vegetation is divided between liquid and solid phases $q_{intr, liq}$ and $q_{intr, ice}$ ($\text{kg m}^{-2} \text{s}^{-1}$)

$$q_{intr, liq} = f_{pi, liq} q_{rain} \quad (2.359)$$

$$q_{intr, ice} = f_{pi, ice} q_{sno} \quad (2.360)$$

where $f_{pi, liq}$ and $f_{pi, ice}$ are the fractions of intercepted precipitation of rain and snow, respectively

$$f_{pi, liq} = \alpha_{liq} \tanh(L + S) \quad (2.361)$$

$$f_{pi, ice} = \alpha_{sno} \{1 - \exp[-0.5(L + S)]\}, \quad (2.362)$$

and L and S are the exposed leaf and stem area index, respectively (section 2.2.1), and the α 's scale the fractional area of a leaf that collects water (Lawrence et al. 2007). Default values of α_{liq} and α_{sno} are set to 1. Throughfall ($\text{kg m}^{-2} \text{s}^{-1}$) is also divided into liquid and solid phases, reaching the ground (soil or snow surface) as

$$q_{thru, liq} = q_{rain} (1 - f_{pi, liq}) \quad (2.363)$$

$$q_{thru, ice} = q_{sno} (1 - f_{pi, ice}) \quad (2.364)$$

Similarly, the liquid and solid canopy drip fluxes are

$$q_{drip, liq} = \frac{W_{can, liq}^{intr} - W_{can, liq}^{max}}{\Delta t} \geq 0 \quad (2.365)$$

$$q_{drip, ice} = \frac{W_{can, sno}^{intr} - W_{can, sno}^{max}}{\Delta t} \geq 0 \quad (2.366)$$

where

$$W_{can, liq}^{intr} = W_{can, liq}^n + q_{intr, liq} \Delta t \geq 0 \quad (2.367)$$

and

$$W_{can, sno}^{intr} = W_{can, sno}^n + q_{intr, ice} \Delta t \geq 0 \quad (2.368)$$

are the canopy liquid water and snow water equivalent after accounting for interception, $W_{can, liq}^n$ and $W_{can, sno}^n$ are the canopy liquid and snow water from the previous time step, and $W_{can, liq}^{max}$ and $W_{can, sno}^{max}$ (kg m^{-2} or mm of H₂O) are the maximum amounts of liquid water and snow the canopy can hold. They are defined by

$$W_{can, liq}^{max} = p_{liq} (L + S) \quad (2.369)$$

$$W_{can, sno}^{max} = p_{ice} (L + S). \quad (2.370)$$

The maximum storage of liquid water is $p_{liq} = 0.1 \text{ kg m}^{-2}$ (*Dickinson et al. 1993*), and that of snow is $p_{sno} = 6 \text{ kg m}^{-2}$, consistent with reported field measurements (*Pomeroy et al. 1998*).

Canopy snow unloading from wind speed u and above-freezing temperatures are modeled from linear fluxes and e-folding times similar to *Roesch et al. (2001)*

$$q_{unl,wind} = \frac{uW_{can,sno}}{1.56 \times 10^5 \text{ m}} \quad (2.371)$$

$$q_{unl,temp} = \frac{W_{can,sno}(T - 270 \text{ K})}{1.87 \times 10^5 \text{ K s}} > 0 \quad (2.372)$$

$$q_{unl,tot} = \min(q_{unl,wind} + q_{unl,temp}, W_{can,sno}) \quad (2.373)$$

The canopy liquid water and snow water equivalent are updated as

$$W_{can,liq}^{n+1} = W_{can,liq}^n + q_{intr,liq} - q_{drip,liq}\Delta t - E_v^{liq}\Delta t \geq 0 \quad (2.374)$$

and

$$W_{can,sno}^{n+1} = W_{can,sno}^n + q_{intr,ice} - (q_{drip,ice} + q_{unl,tot})\Delta t - E_v^{ice}\Delta t \geq 0 \quad (2.375)$$

where E_v^{liq} and E_v^{ice} are partitioned from the stem and leaf surface evaporation E_v^w (Chapter 2.5) based on the vegetation temperature T_v (K) (Chapter 2.5) and its relation to the freezing temperature of water T_f (K) (Table 2.7)

$$E_v^{liq} = \begin{cases} E_v^w & T_v > T_f \\ 0 & T_v \leq T_f \end{cases} \quad (2.376)$$

$$E_v^{ice} = \begin{cases} 0 & T_v > T_f \\ E_v^w & T_v \leq T_f \end{cases}. \quad (2.377)$$

The total rate of liquid and solid precipitation reaching the ground is then

$$q_{grnd,liq} = q_{thru,liq} + q_{drip,liq} \quad (2.378)$$

$$q_{grnd,ice} = q_{thru,ice} + q_{drip,ice} + q_{unl,tot}. \quad (2.379)$$

Solid precipitation reaching the soil or snow surface, $q_{grnd,ice}\Delta t$, is added immediately to the snow pack (Chapter 2.8). The liquid part, $q_{grnd,liq}\Delta t$ is added after surface fluxes (Chapter 2.5) and snow/soil temperatures (Chapter 2.6) have been determined.

The wetted fraction of the canopy (stems plus leaves), which is required for surface flux (Chapter 2.5) calculations, is (*Dickinson et al. 1993*)

$$f_{wet} = \begin{cases} \left[\frac{W_{can}}{p_{liq}(L+S)} \right]^{2/3} \leq 1 & L+S > 0 \\ 0 & L+S = 0 \end{cases} \quad (2.380)$$

while the fraction of the canopy that is dry and transpiring is

$$f_{dry} = \begin{cases} \frac{(1-f_{wet})L}{L+S} & L+S > 0 \\ 0 & L+S = 0 \end{cases}. \quad (2.381)$$

Similarly, the snow-covered fraction of the canopy is used for surface albedo when intercepted snow is present (Chapter 2.3)

$$f_{can,sno} = \begin{cases} \left[\frac{W_{can,sno}}{p_{sno}(L+S)} \right]^{3/20} \leq 1 & L+S > 0 \\ 0 & L+S = 0 \end{cases}. \quad (2.382)$$

2.7.2 Surface Runoff, Surface Water Storage, and Infiltration

The moisture input at the grid cell surface, $q_{liq,0}$, is the sum of liquid precipitation reaching the ground and melt water from snow ($\text{kg m}^{-2} \text{s}^{-1}$). The moisture flux is then partitioned between surface runoff, surface water storage, and infiltration into the soil.

Surface Runoff

The simple TOPMODEL-based (*Beven and Kirkby 1979*) runoff model (SIMTOP) described by *Niu et al. (2005)* is implemented to parameterize runoff. A key concept underlying this approach is that of fractional saturated area f_{sat} , which is determined by the topographic characteristics and soil moisture state of a grid cell. The saturated portion of a grid cell contributes to surface runoff, q_{over} , by the saturation excess mechanism (Dunne runoff)

$$q_{over} = f_{sat} q_{liq,0} \quad (2.383)$$

The fractional saturated area is a function of soil moisture

$$f_{sat} = f_{max} \exp(-0.5 f_{over} z_{\nabla}) \quad (2.384)$$

where f_{max} is the potential or maximum value of f_{sat} , f_{over} is a decay factor (m^{-1}), and z_{∇} is the water table depth (m) (section 2.7.5). The maximum saturated fraction, f_{max} , is defined as the value of the discrete cumulative distribution function (CDF) of the topographic index when the grid cell mean water table depth is zero. Thus, f_{max} is the percent of pixels in a grid cell whose topographic index is larger than or equal to the grid cell mean topographic index. It should be calculated explicitly from the CDF at each grid cell at the resolution that the model is run. However, because this is a computationally intensive task for global applications, f_{max} is calculated once at 0.125° resolution using the 1-km compound topographic indices (CTIs) based on the HYDRO1K dataset (*Verdin and Greenlee 1996*) from USGS following the algorithm in *Niu et al. (2005)* and then area-averaged to the desired model resolution (section 2.2.3). Pixels with CTIs exceeding the 95 percentile threshold in each 0.125° grid cell are excluded from the calculation to eliminate biased estimation of statistics due to large CTI values at pixels on stream networks. For grid cells over regions without CTIs such as Australia, the global mean f_{max} is used to fill the gaps. See *Li et al. (2013b)* for additional details. The decay factor f_{over} for global simulations was determined through sensitivity analysis and comparison with observed runoff to be 0.5 m^{-1} .

Surface Water Storage

A surface water store has been added to the model to represent wetlands and small, sub-grid scale water bodies. As a result, the wetland land unit has been removed. The state variables for surface water are the mass of water W_{sfc} (kg m^{-2}) and temperature T_{h2osfc} (Chapter 2.6). Surface water storage and outflow are functions of fine spatial scale elevation variations called microtopography. The microtopography is assumed to be distributed normally around the grid cell mean elevation. Given the standard deviation of the microtopographic distribution, σ_{micro} (m), the fractional area of the grid cell that is inundated can be calculated. Surface water storage, W_{sfc} , is related to the height (relative to the grid cell mean elevation) of the surface water, d , by

$$W_{sfc} = \frac{d}{2} \left(1 + erf \left(\frac{d}{\sigma_{micro}\sqrt{2}} \right) \right) + \frac{\sigma_{micro}}{\sqrt{2\pi}} e^{\frac{-d^2}{2\sigma_{micro}^2}} \quad (2.385)$$

where erf is the error function. For a given value of W_{sfc} , (2.385) can be solved for d using the Newton-Raphson method. Once d is known, one can determine the fraction of the area that is inundated as

$$f_{h2osfc} = \frac{1}{2} \left(1 + erf \left(\frac{d}{\sigma_{micro}\sqrt{2}} \right) \right) \quad (2.386)$$

No global datasets exist for microtopography, so the default parameterization is a simple function of slope

$$\sigma_{micro} = (\beta + \beta_0)^{\eta} \quad (2.387)$$

where β is the topographic slope, $\beta_0 = (\sigma_{\max})^{\frac{1}{\eta}}$ determines the maximum value of σ_{micro} , and η is an adjustable parameter. Default values in the model are $\sigma_{\max} = 0.4$ and $\eta = -3$.

If the spatial scale of the microtopography is small relative to that of the grid cell, one can assume that the inundated areas are distributed randomly within the grid cell. With this assumption, a result from percolation theory can be used to quantify the fraction of the inundated portion of the grid cell that is interconnected

$$\begin{aligned} f_{\text{connected}} &= (f_{h2osfc} - f_c)^\mu & f_{h2osfc} > f_c \\ f_{\text{connected}} &= 0 & f_{h2osfc} \leq f_c \end{aligned} \quad (2.388)$$

where f_c is a threshold below which no single connected inundated area spans the grid cell and μ is a scaling exponent. Default values of f_c and μ are 0.4 and 0.14, respectively. When the inundated fraction of the grid cell surpasses f_c , the surface water store acts as a linear reservoir

$$q_{out,h2osfc} = k_{h2osfc} f_{\text{connected}} (W_{sfc} - W_c) \frac{1}{\Delta t} \quad (2.389)$$

where $q_{out,h2osfc}$ is the surface water runoff, k_{h2osfc} is a constant, W_c is the amount of surface water present when $f_{h2osfc} = f_c$, and Δt is the model time step. The linear storage coefficient $k_{h2osfc} = \sin(\beta)$ is a function of grid cell mean topographic slope where β is the slope in radians.

Infiltration

The surface moisture flux remaining after surface runoff has been removed,

$$q_{in,surface} = (1 - f_{sat}) q_{liq,0} \quad (2.390)$$

is divided into inputs to surface water ($q_{in,h2osfc}$) and the soil $q_{in,soil}$. If $q_{in,soil}$ exceeds the maximum soil infiltration capacity ($\text{kg m}^{-2} \text{s}^{-1}$),

$$q_{infl,max} = (1 - f_{sat}) \Theta_{ice} k_{sat} \quad (2.391)$$

where Θ_{ice} is an ice impedance factor (section 2.7.3), infiltration excess (Hortonian) runoff is generated

$$q_{infl,excess} = \max(q_{in,soil} - (1 - f_{h2osfc}) q_{infl,max}, 0) \quad (2.392)$$

and transferred from $q_{in,soil}$ to $q_{in,h2osfc}$. After evaporative losses have been removed, these moisture fluxes are

$$q_{in,h2osfc} = f_{h2osfc} q_{in,surface} + q_{infl,excess} - q_{evap,h2osfc} \quad (2.393)$$

and

$$q_{in,soil} = (1 - f_{h2osfc}) q_{in,surface} - q_{infl,excess} - (1 - f_{sno} - f_{h2osfc}) q_{evap,soil}. \quad (2.394)$$

The balance of surface water is then calculated as

$$\Delta W_{sfc} = (q_{in,h2osfc} - q_{out,h2osfc} - q_{drain,h2osfc}) \Delta t. \quad (2.395)$$

Bottom drainage from the surface water store

$$q_{drain,h2osfc} = \min \left(f_{h2osfc} q_{infl,max}, \frac{W_{sfc}}{\Delta t} \right) \quad (2.396)$$

is then added to $q_{in,soil}$ giving the total infiltration into the surface soil layer

$$q_{infl} = q_{in,soil} + q_{drain,h2osfc} \quad (2.397)$$

Infiltration q_{infl} and explicit surface runoff q_{over} are not allowed for glaciers.

2.7.3 Soil Water

Soil water is predicted from a multi-layer model, in which the vertical soil moisture transport is governed by infiltration, surface and sub-surface runoff, gradient diffusion, gravity, and canopy transpiration through root extraction (Figure 2.7).

For one-dimensional vertical water flow in soils, the conservation of mass is stated as

$$\frac{\partial \theta}{\partial t} = -\frac{\partial q}{\partial z} - e \quad (2.398)$$

where θ is the volumetric soil water content (mm^3 of water / mm^{-3} of soil), t is time (s), z is height above some datum in the soil column (mm) (positive upwards), q is soil water flux ($\text{kg m}^{-2} \text{s}^{-1}$ or mm s^{-1}) (positive upwards), and e is a soil moisture sink term ($\text{mm of water mm}^{-1}$ of soil s^{-1}) (ET loss). This equation is solved numerically by dividing the soil column into multiple layers in the vertical and integrating downward over each layer with an upper boundary condition of the infiltration flux into the top soil layer q_{infl} and a zero-flux lower boundary condition at the bottom of the soil column (sub-surface runoff is removed later in the timestep, section 2.7.5).

The soil water flux q in equation can be described by Darcy's law (Dingman 2002)

$$q = -k \frac{\partial \psi_h}{\partial z} \quad (2.399)$$

where k is the hydraulic conductivity (mm s^{-1}), and ψ_h is the hydraulic potential (mm). The hydraulic potential is

$$\psi_h = \psi_m + \psi_z \quad (2.400)$$

where ψ_m is the soil matric potential (mm) (which is related to the adsorptive and capillary forces within the soil matrix), and ψ_z is the gravitational potential (mm) (the vertical distance from an arbitrary reference elevation to a point in the soil). If the reference elevation is the soil surface, then $\psi_z = z$. Letting $\psi = \psi_m$, Darcy's law becomes

$$q = -k \left[\frac{\partial (\psi + z)}{\partial z} \right]. \quad (2.401)$$

Equation (2.401) can be further manipulated to yield

$$q = -k \left[\frac{\partial (\psi + z)}{\partial z} \right] = -k \left(\frac{\partial \psi}{\partial z} + 1 \right). \quad (2.402)$$

Substitution of this equation into equation (2.398), with $e = 0$, yields the Richards equation (Dingman 2002)

$$\frac{\partial \theta}{\partial t} = \frac{\partial}{\partial z} \left[k \left(\frac{\partial \psi}{\partial z} + 1 \right) \right]. \quad (2.403)$$

In practice (Section 2.7.3), changes in soil water content are predicted from (2.398) using finite-difference approximations for (2.403).

Hydraulic Properties

The hydraulic conductivity k_i (mm s^{-1}) and the soil matric potential ψ_i (mm) for layer i vary with volumetric soil water θ_i and soil texture. As with the soil thermal properties (section 2.6.3) the hydraulic properties of the soil are assumed to be a weighted combination of the mineral properties, which are determined according to sand and clay contents based on work by Clapp and Hornberger (1978) and Cosby et al. (1984), and organic properties of the soil (Lawrence and Slater 2008).

The hydraulic conductivity is defined at the depth of the interface of two adjacent layers $z_{h,i}$ (Figure 2.8) and is a function of the saturated hydraulic conductivity $k_{sat}[z_{h,i}]$, the liquid volumetric soil moisture of the two layers θ_i and

θ_{i+1} and an ice impedance factor Θ_{ice}

$$k[z_h, i] = \begin{cases} \Theta_{ice} k_{sat}[z_h, i] \left[\frac{0.5(\theta_i + \theta_{i+1})}{0.5(\theta_{sat,i} + \theta_{sat,i+1})} \right]^{2B_i+3} & 1 \leq i \leq N_{levsoi} - 1 \\ \Theta_{ice} k_{sat}[z_h, i] \left(\frac{\theta_i}{\theta_{sat,i}} \right)^{2B_i+3} & i = N_{levsoi} \end{cases}. \quad (2.404)$$

The ice impedance factor is a function of ice content, and is meant to quantify the increased tortuosity of the water flow when part of the pore space is filled with ice. [Swenson et al. \(2012\)](#) used a power law form

$$\Theta_{ice} = 10^{-\Omega F_{ice}} \quad (2.405)$$

where $\Omega = 6$ and $F_{ice} = \frac{\theta_{ice}}{\theta_{sat}}$ is the ice-filled fraction of the pore space.

Because the hydraulic properties of mineral and organic soil may differ significantly, the bulk hydraulic properties of each soil layer are computed as weighted averages of the properties of the mineral and organic components. The water content at saturation (i.e. porosity) is

$$\theta_{sat,i} = (1 - f_{om,i})\theta_{sat,min,i} + f_{om,i}\theta_{sat,om} \quad (2.406)$$

where $f_{om,i}$ is the soil organic matter fraction, $\theta_{sat,om} = 0.9$ ([Farouki 1981](#); [Letts et al. 2000](#)) is the porosity of organic matter and the porosity of the mineral soil $\theta_{sat,min,i}$ is

$$\theta_{sat,min,i} = 0.489 - 0.00126(\%sand)_i. \quad (2.407)$$

The exponent B_i is

$$B_i = (1 - f_{om,i})B_{min,i} + f_{om,i}B_{om} \quad (2.408)$$

where $B_{om} = 2.7$ ([Letts et al. 2000](#)) and

$$B_{min,i} = 2.91 + 0.159(\%clay)_i. \quad (2.409)$$

The soil matric potential (mm) is defined at the node depth z_i of each layer i (Figure 2.8)

$$\psi_i = \psi_{sat,i} \left(\frac{\theta_i}{\theta_{sat,i}} \right)^{-B_i} \geq -1 \times 10^8 \quad 0.01 \leq \frac{\theta_i}{\theta_{sat,i}} \leq 1 \quad (2.410)$$

where the saturated soil matric potential (mm) is

$$\psi_{sat,i} = (1 - f_{om,i})\psi_{sat,min,i} + f_{om,i}\psi_{sat,om} \quad (2.411)$$

where $\psi_{sat,om} = -10.3$ mm ([Letts et al. 2000](#)) is the saturated organic matter matric potential and the saturated mineral soil matric potential $\psi_{sat,min,i}$ is

$$\psi_{sat,min,i} = -10.0 \times 10^{1.88 - 0.0131(\%sand)_i}. \quad (2.412)$$

The saturated hydraulic conductivity, $k_{sat}[z_h, i]$ (mm s⁻¹), for organic soils ($k_{sat,om}$) may be two to three orders of magnitude larger than that of mineral soils ($k_{sat,min}$). Bulk soil layer values of k_{sat} calculated as weighted averages based on f_{om} may therefore be determined primarily by the organic soil properties even for values of f_{om} as low as 1 %. To better represent the influence of organic soil material on the grid cell average saturated hydraulic conductivity, the soil organic matter fraction is further subdivided into “connected” and “unconnected” fractions using a result from percolation theory ([Stauffer and Aharony 1994](#), [Berkowitz and Balberg 1992](#)). Assuming that the organic and mineral fractions are randomly distributed throughout a soil layer, percolation theory predicts that above a threshold value $f_{om} = f_{threshold}$, connected flow pathways consisting of organic material only exist and span the soil space. Flow through these pathways interacts only with organic material, and thus can be described by $k_{sat,om}$. This fraction of the grid cell is given by

$$\begin{aligned} f_{perc} &= N_{perc} (f_{om} - f_{threshold})^{\beta_{perc}} f_{om} & f_{om} \geq f_{threshold} \\ f_{perc} &= 0 & f_{om} < f_{threshold} \end{aligned} \quad (2.413)$$

where $\beta^{perc} = 0.139$, $f_{threshold} = 0.5$, and $N_{perc} = (1 - f_{threshold})^{-\beta_{perc}}$. In the unconnected portion of the grid cell, $f_{uncon} = (1 - f_{perc})$, the saturated hydraulic conductivity is assumed to correspond to flow pathways that pass through the mineral and organic components in series

$$k_{sat, uncon} = f_{uncon} \left(\frac{(1 - f_{om})}{k_{sat, min}} + \frac{(f_{om} - f_{perc})}{k_{sat, om}} \right)^{-1}. \quad (2.414)$$

where saturated hydraulic conductivity for mineral soil depends on soil texture (*Cosby et al. 1984*) as

$$k_{sat, min} [z_h, i] = 0.0070556 \times 10^{-0.884+0.0153(\%sand)_i}. \quad (2.415)$$

The bulk soil layer saturated hydraulic conductivity is then computed as

$$k_{sat} [z_h, i] = f_{uncon, i} k_{sat, uncon} [z_h, i] + (1 - f_{uncon, i}) k_{sat, om} [z_h, i]. \quad (2.416)$$

Numerical Solution

With reference to Figure 2.8, the equation for conservation of mass (equation (2.398)) can be integrated over each layer as

$$\int_{-z_h, i}^{-z_h, i-1} \frac{\partial \theta}{\partial t} dz = - \int_{-z_h, i}^{-z_h, i-1} \frac{\partial q}{\partial z} dz - \int_{-z_h, i}^{-z_h, i-1} e dz. \quad (2.417)$$

Note that the integration limits are negative since z is defined as positive upward from the soil surface. This equation can be written as

$$\Delta z_i \frac{\partial \theta_{liq, i}}{\partial t} = -q_{i-1} + q_i - e_i \quad (2.418)$$

where q_i is the flux of water across interface z_h, i , q_{i-1} is the flux of water across interface $z_h, i-1$, and e_i is a layer-averaged soil moisture sink term (ET loss) defined as positive for flow out of the layer (mm s^{-1}). Taking the finite difference with time and evaluating the fluxes implicitly at time $n + 1$ yields

$$\frac{\Delta z_i \Delta \theta_{liq, i}}{\Delta t} = -q_{i-1}^{n+1} + q_i^{n+1} - e_i \quad (2.419)$$

where $\Delta \theta_{liq, i} = \theta_{liq, i}^{n+1} - \theta_{liq, i}^n$ is the change in volumetric soil liquid water of layer i in time Δt and Δz_i is the thickness of layer i (mm).

The water removed by transpiration in each layer e_i is a function of the total transpiration E_v^t (Chapter 2.5) and the effective root fraction $r_{e, i}$

$$e_i = r_{e, i} E_v^t. \quad (2.420)$$

Shown are three soil layers, $i - 1$, i , and $i + 1$. The soil matric potential ψ and volumetric soil water θ_{liq} are defined at the layer node depth z . The hydraulic conductivity $k[z_h]$ is defined at the interface of two layers z_h . The layer thickness is Δz . The soil water fluxes q_{i-1} and q_i are defined as positive upwards. The soil moisture sink term e (ET loss) is defined as positive for flow out of the layer.

Note that because more than one plant functional type (PFT) may share a soil column, the transpiration E_v^t is a weighted sum of transpiration from all PFTs whose weighting depends on PFT area as

$$E_v^t = \sum_{j=1}^{npft} (E_v^t)_j (wt)_j \quad (2.421)$$

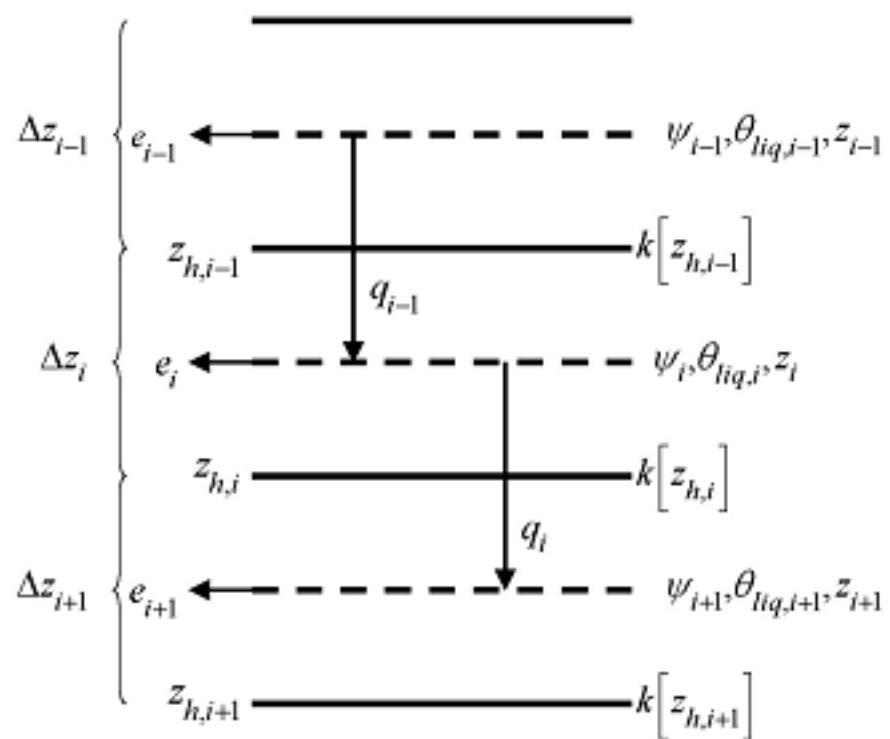


Figure 2.8: Schematic diagram of numerical scheme used to solve for soil water fluxes.

where $npft$ is the number of PFTs sharing a soil column, $(E_v^t)_j$ is the transpiration from the j^{th} PFT on the column, and $(wt)_j$ is the relative area of the j^{th} PFT with respect to the column. The effective root fraction $r_{e,i}$ is also a column-level quantity that is a weighted sum over all PFTs. The weighting depends on the per unit area transpiration of each PFT and its relative area as

$$r_{e,i} = \frac{\sum_{j=1}^{npft} (r_{e,i})_j (E_v^t)_j (wt)_j}{\sum_{j=1}^{npft} (E_v^t)_j (wt)_j} \quad (2.422)$$

where $(r_{e,i})_j$ is the effective root fraction for the j^{th} PFT

$$\begin{aligned} (r_{e,i})_j &= \frac{(r_i)_j (w_i)_j}{(\beta_t)_j} & (\beta_t)_j > 0 \\ (r_{e,i})_j &= 0 & (\beta_t)_j = 0 \end{aligned} \quad (2.423)$$

and $(r_i)_j$ is the fraction of roots in layer i (Chapter 2.9), $(w_i)_j$ is a soil dryness or plant wilting factor for layer i (Chapter 2.9), and $(\beta_t)_j$ is a wetness factor for the total soil column for the j^{th} PFT (Chapter 2.9).

The soil water fluxes in (2.419), which are a function of $\theta_{liq,i}$ and $\theta_{liq,i+1}$ because of their dependence on hydraulic conductivity and soil matric potential, can be linearized about θ using a Taylor series expansion as

$$q_i^{n+1} = q_i^n + \frac{\partial q_i}{\partial \theta_{liq,i}} \Delta \theta_{liq,i} + \frac{\partial q_i}{\partial \theta_{liq,i+1}} \Delta \theta_{liq,i+1} \quad (2.424)$$

$$q_{i-1}^{n+1} = q_{i-1}^n + \frac{\partial q_{i-1}}{\partial \theta_{liq,i-1}} \Delta \theta_{liq,i-1} + \frac{\partial q_{i-1}}{\partial \theta_{liq,i}} \Delta \theta_{liq,i}. \quad (2.425)$$

Substitution of these expressions for q_i^{n+1} and q_{i-1}^{n+1} into (2.419) results in a general tridiagonal equation set of the form

$$r_i = a_i \Delta \theta_{liq,i-1} + b_i \Delta \theta_{liq,i} + c_i \Delta \theta_{liq,i+1} \quad (2.426)$$

where

$$a_i = -\frac{\partial q_{i-1}}{\partial \theta_{liq,i-1}} \quad (2.427)$$

$$b_i = \frac{\partial q_i}{\partial \theta_{liq,i}} - \frac{\partial q_{i-1}}{\partial \theta_{liq,i}} - \frac{\Delta z_i}{\Delta t} \quad (2.428)$$

$$c_i = \frac{\partial q_i}{\partial \theta_{liq,i+1}} \quad (2.429)$$

$$r_i = q_{i-1}^n - q_i^n + e_i. \quad (2.430)$$

The tridiagonal equation set is solved over $i = 1, \dots, N_{levsoi}$.

The finite-difference forms of the fluxes and partial derivatives in equations (2.427) - (2.430) can be obtained from equation as

$$q_{i-1}^n = -k [z_{h,i-1}] \left[\frac{(\psi_{i-1} - \psi_i) + (z_i - z_{i-1})}{z_i - z_{i-1}} \right] \quad (2.431)$$

$$q_i^n = -k [z_{h,i}] \left[\frac{(\psi_i - \psi_{i+1}) + (z_{i+1} - z_i)}{z_{i+1} - z_i} \right] \quad (2.432)$$

$$\frac{\partial q_{i-1}}{\partial \theta_{liq,i-1}} = - \left[\frac{k [z_{h,i-1}]}{z_i - z_{i-1}} \frac{\partial \psi_{i-1}}{\partial \theta_{liq,i-1}} \right] - \frac{\partial k [z_{h,i-1}]}{\partial \theta_{liq,i-1}} \left[\frac{(\psi_{i-1} - \psi_i) + (z_i - z_{i-1})}{z_i - z_{i-1}} \right] \quad (2.433)$$

$$\frac{\partial q_{i-1}}{\partial \theta_{liq,i}} = \left[\frac{k [z_{h,i-1}]}{z_i - z_{i-1}} \frac{\partial \psi_i}{\partial \theta_{liq,i}} \right] - \frac{\partial k [z_{h,i-1}]}{\partial \theta_{liq,i}} \left[\frac{(\psi_{i-1} - \psi_i) + (z_i - z_{i-1})}{z_i - z_{i-1}} \right] \quad (2.434)$$

$$\frac{\partial q_i}{\partial \theta_{liq,i}} = - \left[\frac{k [z_{h,i}]}{z_{i+1} - z_i} \frac{\partial \psi_i}{\partial \theta_{liq,i}} \right] - \frac{\partial k [z_{h,i}]}{\partial \theta_{liq,i}} \left[\frac{(\psi_i - \psi_{i+1}) + (z_{i+1} - z_i)}{z_{i+1} - z_i} \right] \quad (2.435)$$

$$\frac{\partial q_i}{\partial \theta_{liq,i+1}} = \left[\frac{k [z_{h,i}]}{z_{i+1} - z_i} \frac{\partial \psi_{i+1}}{\partial \theta_{liq,i+1}} \right] - \frac{\partial k [z_{h,i}]}{\partial \theta_{liq,i+1}} \left[\frac{(\psi_i - \psi_{i+1}) + (z_{i+1} - z_i)}{z_{i+1} - z_i} \right]. \quad (2.436)$$

The derivatives of the soil matric potential at the node depth are derived from (2.410)

$$\frac{\partial \psi_{i-1}}{\partial \theta_{liq,i-1}} = -B_{i-1} \frac{\psi_{i-1}}{\theta_{i-1}} \quad (2.437)$$

$$\frac{\partial \psi_i}{\partial \theta_{liq,i}} = -B_i \frac{\psi_i}{\theta_i} \quad (2.438)$$

$$\frac{\partial \psi_{i+1}}{\partial \theta_{liq,i+1}} = -B_{i+1} \frac{\psi_{i+1}}{\theta_{i+1}} \quad (2.439)$$

with the constraint $0.01 \theta_{sat,i} \leq \theta_i \leq \theta_{sat,i}$.

The derivatives of the hydraulic conductivity at the layer interface are derived from (2.404)

$$\frac{\partial k [z_{h,i-1}]}{\partial \theta_{liq,i-1}} = \frac{\partial k [z_{h,i-1}]}{\partial \theta_{liq,i}} = (2B_{i-1} + 3) \bar{\Theta}_{ice} k_{sat} [z_{h,i-1}] \left[\frac{\bar{\theta}_{liq}}{\bar{\theta}_{sat}} \right]^{2B_{i-1}+2} \left(\frac{0.5}{\bar{\theta}_{sat}} \right) \quad (2.440)$$

where $\bar{\Theta}_{ice} = \Theta(\bar{\theta}_{ice})$ (2.405), $\bar{\theta}_{ice} = 0.5(\theta_{ice,i-1} + \theta_{ice,i})$, $\bar{\theta}_{liq} = 0.5(\theta_{liq,i-1} + \theta_{liq,i})$, and $\bar{\theta}_{sat} = 0.5(\theta_{sat,i-1} + \theta_{sat,i})$

and

$$\frac{\partial k [z_{h,i}]}{\partial \theta_{liq,i}} = \frac{\partial k [z_{h,i}]}{\partial \theta_{liq,i+1}} = (2B_i + 3) \bar{\Theta}_{ice} k_{sat} [z_{h,i}] \left[\frac{\bar{\theta}_{liq}}{\bar{\theta}_{sat}} \right]^{2B_i+2} \left(\frac{0.5}{\bar{\theta}_{sat}} \right). \quad (2.441)$$

where $\bar{\theta}_{liq} = 0.5(\theta_i + \theta_{i+1})$, $\bar{\theta}_{sat} = 0.5(\theta_{sat,i} + \theta_{sat,i+1})$.

Equation set for layer $i = 1$

For the top soil layer ($i = 1$), the boundary condition is the infiltration rate (section 2.7.2), $q_{i-1}^{n+1} = -q_{infl}^{n+1}$, and the water balance equation is

$$\frac{\Delta z_i \Delta \theta_{liq,i}}{\Delta t} = q_{infl}^{n+1} + q_i^{n+1} - e_i. \quad (2.442)$$

After grouping like terms, the coefficients of the tridiagonal set of equations for $i = 1$ are

$$a_i = 0 \quad (2.443)$$

$$b_i = \frac{\partial q_i}{\partial \theta_{liq, i}} - \frac{\Delta z_i}{\Delta t} \quad (2.444)$$

$$c_i = \frac{\partial q_i}{\partial \theta_{liq, i+1}} \quad (2.445)$$

$$r_i = q_{infl}^{n+1} - q_i^n + e_i. \quad (2.446)$$

Equation set for layers $i = 2, \dots, N_{levsoi} - 1$

The coefficients of the tridiagonal set of equations for $i = 2, \dots, N_{levsoi} - 1$ are

$$a_i = -\frac{\partial q_{i-1}}{\partial \theta_{liq, i-1}} \quad (2.447)$$

$$b_i = \frac{\partial q_i}{\partial \theta_{liq, i}} - \frac{\partial q_{i-1}}{\partial \theta_{liq, i}} - \frac{\Delta z_i}{\Delta t} \quad (2.448)$$

$$c_i = \frac{\partial q_i}{\partial \theta_{liq, i+1}} \quad (2.449)$$

$$r_i = q_{i-1}^n - q_i^n + e_i. \quad (2.450)$$

Equation set for layer $i = N_{levsoi}$

For the lowest soil layer ($i = N_{levsoi}$), a zero-flux bottom boundary condition is applied ($q_i^n = 0$) and the coefficients of the tridiagonal set of equations for $i = N_{levsoi}$ are

$$a_i = -\frac{\partial q_{i-1}}{\partial \theta_{liq, i-1}} \quad (2.451)$$

$$b_i = \frac{\partial q_i}{\partial \theta_{liq, i}} - \frac{\partial q_{i-1}}{\partial \theta_{liq, i}} - \frac{\Delta z_i}{\Delta t} \quad (2.452)$$

$$c_i = 0 \quad (2.453)$$

$$r_i = q_{i-1}^n + e_i. \quad (2.454)$$

Adaptive Time Stepping

The length of the time step is adjusted in order to improve the accuracy and stability of the numerical solutions. The difference between two numerical approximations is used to estimate the temporal truncation error, and then the step size Δt_{sub} is adjusted to meet a user-prescribed error tolerance [Kavetski et al., 2002]. The temporal truncation error is estimated by comparing the flux obtained from the first-order Taylor series expansion (q_{i-1}^{n+1} and q_i^{n+1} , equations (2.424) and (2.425)) against the flux at the start of the time step (q_{i-1}^n and q_i^n). Since the tridiagonal solution already provides an estimate of $\Delta\theta_{liq,i}$, it is convenient to compute the error for each of the i layers from equation (2.419) as

$$\epsilon_i = \left[\frac{\Delta\theta_{liq,i} \Delta z_i}{\Delta t_{sub}} - (q_{i-1}^n - q_i^n + e_i) \right] \frac{\Delta t_{sub}}{2} \quad (2.455)$$

and the maximum absolute error across all layers as

$$\epsilon_{crit} = \max(|\epsilon_i|) \quad 1 \leq i \leq nlevsoi . \quad (2.456)$$

The adaptive step size selection is based on specified upper and lower error tolerances, τ_U and τ_L . The solution is accepted if $\epsilon_{crit} \leq \tau_U$ and the procedure repeats until the adaptive sub-stepping spans the full model time step (the sub-steps are doubled if $\epsilon_{crit} \leq \tau_L$, i.e., if the solution is very accurate). Conversely, the solution is rejected if $\epsilon_{crit} > \tau_U$. In this case the length of the sub-steps is halved and a new solution is obtained. The halving of substeps continues until either $\epsilon_{crit} \leq \tau_U$ or the specified minimum time step length is reached.

Upon solution of the tridiagonal equation set, the liquid water contents are updated as follows

$$w_{liq,i}^{n+1} = w_{liq,i}^n + \Delta\theta_{liq,i} \Delta z_i \quad i = 1, \dots, N_{levsoi}. \quad (2.457)$$

The volumetric water content is

$$\theta_i = \frac{w_{liq,i}}{\Delta z_i \rho_{liq}} + \frac{w_{ice,i}}{\Delta z_i \rho_{ice}}. \quad (2.458)$$

2.7.4 Frozen Soils and Perched Water Table

When soils freeze, the power-law form of the ice impedance factor (section 2.7.3) can greatly decrease the hydraulic conductivity of the soil, leading to nearly impermeable soil layers. When unfrozen soil layers are present above relatively ice-rich frozen layers, the possibility exists for perched saturated zones. Lateral drainage from perched saturated regions is parameterized as a function of the thickness of the saturated zone

$$q_{drai,perch} = k_{drai,perch} (z_{frost} - z_{\nabla,perch}) \quad (2.459)$$

where $k_{drai,perch}$ depends on topographic slope and soil hydraulic conductivity,

$$k_{drai,perch} = 10^{-5} \sin(\beta) \left(\frac{\sum_{i=N_{perch}}^{i=N_{frost}} \Theta_{ice,i} k_{sat}[z_i] \Delta z_i}{\sum_{i=N_{perch}}^{i=N_{frost}} \Delta z_i} \right) \quad (2.460)$$

where Θ_{ice} is an ice impedance factor, β is the mean grid cell topographic slope in radians, z_{frost} is the depth to the frost table, and $z_{\nabla,perch}$ is the depth to the perched saturated zone. The frost table z_{frost} is defined as the shallowest frozen layer having an unfrozen layer above it, while the perched water table $z_{\nabla,perch}$ is defined as the depth at which the volumetric water content drops below a specified threshold. The default threshold is set to 0.9. Drainage from the perched saturated zone $q_{drai,perch}$ is removed from layers N_{perch} through N_{frost} , which are the layers containing $z_{\nabla,perch}$ and, z_{frost} respectively.

2.7.5 Lateral Sub-surface Runoff

Lateral sub-surface runoff occurs when saturated soil moisture conditions exist within the soil column. Sub-surface runoff is

$$q_{drai} = \Theta_{ice} K_{baseflow} \tan(\beta) \Delta z_{sat}^{N_{baseflow}}, \quad (2.461)$$

where $K_{baseflow}$ is a calibration parameter, β is the topographic slope, the exponent $N_{baseflow} = 1$, and Δz_{sat} is the thickness of the saturated portion of the soil column.

The saturated thickness is

$$\Delta z_{sat} = z_{bedrock} - z_\nabla, \quad (2.462)$$

where the water table z_∇ is determined by finding the first soil layer above the bedrock depth (section 2.2.2) in which the volumetric water content drops below a specified threshold. The default threshold is set to 0.9.

The specific yield, S_y , which depends on the soil properties and the water table location, is derived by taking the difference between two equilibrium soil moisture profiles whose water tables differ by an infinitesimal amount

$$S_y = \theta_{sat} \left(1 - \left(1 + \frac{z_\nabla}{\Psi_{sat}} \right)^{\frac{-1}{B}} \right) \quad (2.463)$$

where B is the Clapp-Hornberger exponent. Because S_y is a function of the soil properties, it results in water table dynamics that are consistent with the soil water fluxes described in section 2.7.3.

After the above calculations, two numerical adjustments are implemented to keep the liquid water content of each soil layer ($w_{liq,i}$) within physical constraints of $w_{liq}^{\min} \leq w_{liq,i} \leq (\theta_{sat,i} - \theta_{ice,i}) \Delta z_i$ where $w_{liq}^{\min} = 0.01$ (mm). First, beginning with the bottom soil layer $i = N_{levsoi}$, any excess liquid water in each soil layer ($w_{liq,i}^{excess} = w_{liq,i} - (\theta_{sat,i} - \theta_{ice,i}) \Delta z_i \geq 0$) is successively added to the layer above. Any excess liquid water that remains after saturating the entire soil column (plus a maximum surface ponding depth $w_{liq}^{pond} = 10$ kg m⁻²), is added to drainage q_{drai} . Second, to prevent negative $w_{liq,i}$, each layer is successively brought up to $w_{liq,i} = w_{liq}^{\min}$ by taking the required amount of water from the layer below. If this results in $w_{liq,N_{levsoi}} < w_{liq}^{\min}$, then the layers above are searched in succession for the required amount of water ($w_{liq}^{\min} - w_{liq,N_{levsoi}}$) and removed from those layers subject to the constraint $w_{liq,i} \geq w_{liq}^{\min}$. If sufficient water is not found, then the water is removed from W_t and q_{drai} .

The soil surface layer liquid water and ice contents are then updated for dew q_{sdew} , frost q_{frost} , or sublimation q_{subl} (section 2.5.4) as

$$w_{liq,1}^{n+1} = w_{liq,1}^n + q_{sdew} \Delta t \quad (2.464)$$

$$w_{ice,1}^{n+1} = w_{ice,1}^n + q_{frost} \Delta t \quad (2.465)$$

$$w_{ice,1}^{n+1} = w_{ice,1}^n - q_{subl} \Delta t. \quad (2.466)$$

Sublimation of ice is limited to the amount of ice available.

2.7.6 Runoff from glaciers and snow-capped surfaces

All surfaces are constrained to have a snow water equivalent $W_{sno} \leq 1000$ kg m⁻². For snow-capped surfaces, the solid and liquid precipitation reaching the snow surface and dew in solid or liquid form, is separated into solid $q_{snwcp,ice}$ and liquid $q_{snwcp,liq}$ runoff terms

$$q_{snwcp,ice} = q_{grnd,ice} + q_{frost} \quad (2.467)$$

$$q_{snwcp,liq} = q_{grnd,liq} + q_{dew} \quad (2.468)$$

and snow pack properties are unchanged. The $q_{snwcp,ice}$ runoff is sent to the River Transport Model (RTM) (Chapter 11) where it is routed to the ocean as an ice stream and, if applicable, the ice is melted there.

For snow-capped surfaces other than glaciers and lakes the $q_{snwcp,liq}$ runoff is assigned to the glaciers and lakes runoff term q_{rgwl} (e.g. $q_{rgwl} = q_{snwcp,liq}$). For glacier surfaces the runoff term q_{rgwl} is calculated from the residual of the water balance

$$q_{rgwl} = q_{grnd,ice} + q_{grnd,liq} - E_g - E_v - \frac{(W_b^{n+1} - W_b^n)}{\Delta t} - q_{snwcp,ice} \quad (2.469)$$

where W_b^n and W_b^{n+1} are the water balances at the beginning and ending of the time step defined as

$$W_b = W_{can} + W_{sno} + \sum_{i=1}^N (w_{ice,i} + w_{liq,i}). \quad (2.470)$$

Currently, glaciers are non-vegetated and $E_v = W_{can} = 0$. The contribution of lake runoff to q_{rgwl} is described in section 2.12.6. The runoff term q_{rgwl} may be negative for glaciers and lakes, which reduces the total amount of runoff available to the river routing model (Chapter 2.14).

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2.8 Snow Hydrology

The parameterizations for snow are based primarily on [Anderson \(1976\)](#), [Jordan \(1991\)](#), and [Dai and Zeng \(1997\)](#). The snowpack can have up to five layers. These layers are indexed in the Fortran code as $i = -4, -3, -2, -1, 0$ where layer $i = 0$ is the snow layer next to the top soil layer and layer $i = -4$ is the top layer of a five-layer snow pack. Since the number of snow layers varies according to the snow depth, we use the notation $snl + 1$ to describe the top layer of snow for the variable layer snow pack, where snl is the negative of the number of snow layers. Refer to [Figure 2.9](#) for an example of the snow layer structure for a three layer snow pack.

Shown are three snow layers, $i = -2$, $i = -1$, and $i = 0$. The layer node depth is z , the layer interface is z_h , and the layer thickness is Δz .

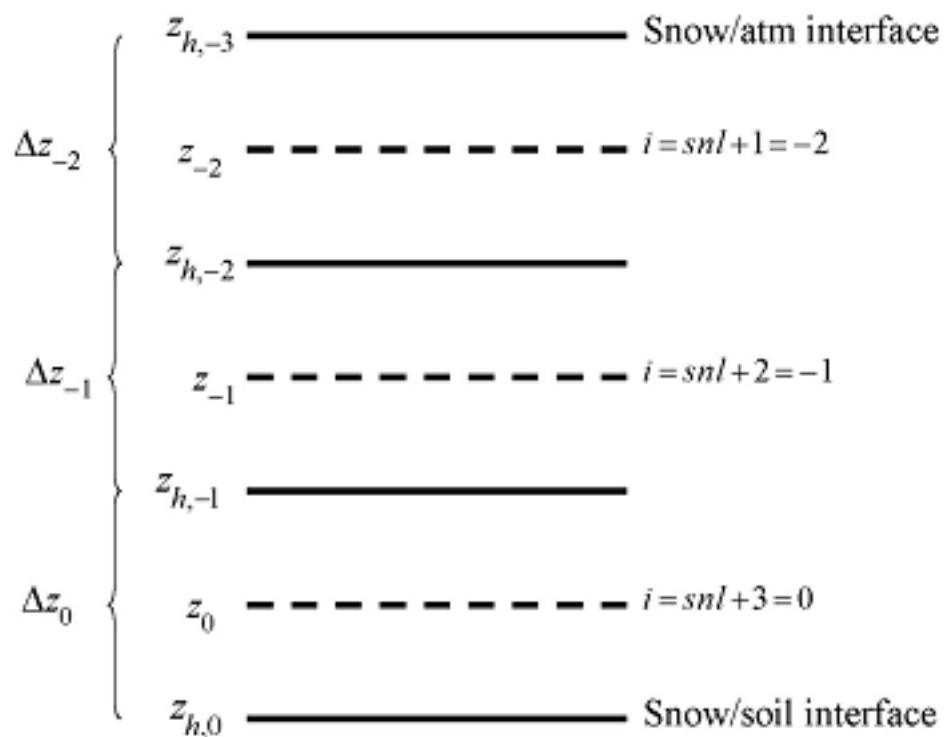
The state variables for snow are the mass of water $w_{liq,i}$ (kg m^{-2}), mass of ice $w_{ice,i}$ (kg m^{-2}), layer thickness Δz_i (m), and temperature T_i (Chapter 2.6). The water vapor phase is neglected. Snow can also exist in the model without being represented by explicit snow layers. This occurs when the snowpack is less than a specified minimum snow depth ($z_{sno} < 0.01$ m). In this case, the state variable is the mass of snow W_{sno} (kg m^{-2}).

[Section 2.8.1](#) describes the calculation of fractional snow covered area, which is used in the surface albedo calculation (Chapter 2.3) and the surface flux calculations (Chapter 2.5). The following two sections ([2.8.2](#) and [2.8.3](#)) describe the ice and water content of the snow pack assuming that at least one snow layer exists. [Section 2.8.4](#) describes how black and organic carbon and mineral dust particles are represented within snow, including meltwater flushing. See [Section 2.8.5](#) for a description of how a snow layer is initialized.

2.8.1 Snow Covered Area Fraction

The fraction of the ground covered by snow, f_{sno} , is based on the method of [Swenson and Lawrence \(2012\)](#). Because the processes governing snowfall and snowmelt differ, changes in f_{sno} are calculated separately for accumulation and depletion. When snowfall occurs, f_{sno} is updated as

$$f_{sno}^{n+1} = 1 - ((1 - \tanh(k_{accum} q_{sno} \Delta t)) (1 - f_{sno}^n)) \quad (2.471)$$

Figure 2.9: Example of three layer snow pack ($snl = -3$).

where k_{accum} is a constant whose default value is 0.1, $q_{sno}\Delta t$ is the amount of new snow, f_{sno}^{n+1} is the updated snow covered fraction (SCF), and f_{sno}^n is the SCF from the previous time step.

When snow melt occurs, f_{sno} is calculated from the depletion curve

$$f_{sno} = 1 - \left(\frac{\cos^{-1}(2R_{sno} - 1)}{\pi} \right)^{N_{melt}} \quad (2.472)$$

where R_{sno} is the ratio of W_{sno} to the maximum accumulated snow W_{max} , and N_{melt} is a parameter that depends on the topographic variability within the grid cell. Whenever W_{sno} reaches zero, W_{max} is reset to zero. The depletion curve shape parameter is defined as

$$N_{melt} = \frac{200}{\min(10, \sigma_{topo})} \quad (2.473)$$

The standard deviation of the elevation within a grid cell, σ_{topo} , is calculated from a high resolution DEM (a 1km DEM is used for CLM).

2.8.2 Ice Content

The conservation equation for mass of ice in snow layers is

$$\frac{\partial w_{ice, i}}{\partial t} = \begin{cases} f_{sno} q_{ice, i-1} - \frac{(\Delta w_{ice, i})_p}{\Delta t} & i = snl + 1 \\ -\frac{(\Delta w_{ice, i})_p}{\Delta t} & i = snl + 2, \dots, 0 \end{cases} \quad (2.474)$$

where $q_{ice, i-1}$ is the rate of ice accumulation from precipitation or frost or the rate of ice loss from sublimation ($\text{kg m}^{-2} \text{s}^{-1}$) in the top layer and $(\Delta w_{ice, i})_p / \Delta t$ is the change in ice due to phase change (melting rate) (section 2.6.2). The term $q_{ice, i-1}$ is computed in two steps as

$$q_{ice, i-1} = q_{grnd, ice} + (q_{frost} - q_{subl}) \quad (2.475)$$

where $q_{grnd, ice}$ is the rate of solid precipitation reaching the ground (section 2.7.1) and q_{frost} and q_{subl} are gains due to frost and losses due to sublimation, respectively (section 2.5.4). In the first step, immediately after $q_{grnd, ice}$ has been determined after accounting for interception (section 2.7.1), a new snow depth z_{sno} (m) is calculated from

$$z_{sno}^{n+1} = z_{sno}^n + \Delta z_{sno} \quad (2.476)$$

where

$$\Delta z_{sno} = \frac{q_{grnd, ice} \Delta t}{f_{sno} \rho_{sno}} \quad (2.477)$$

and ρ_{sno} is the bulk density of newly fallen snow (kg m^{-3}) (van Kampenhout et al. (2017), Anderson (1976))

$$\rho_{sno} = \begin{cases} 50 + 1.7(17)^{1.5} & T_{atm} > T_f + 2 \\ 50 + 1.7(T_{atm} - T_f + 15)^{1.5} & T_f - 15 < T_{atm} \leq T_f + 2 \\ -3.833(T_{atm} - T_f) - 0.0333(T_{atm} - T_f)^2 & T_{atm} \leq T_f - 15 \end{cases} \quad (2.478)$$

where T_{atm} is the atmospheric temperature (K), and T_f is the freezing temperature of water (K) (Table 2.7). When wind speed W_{atm} is greater than 0.1 m s^{-1} , snow density increases due to wind-driven compaction according to (van Kampenhout et al. 2017)

$$\Delta \rho_{sno} = 266.861 \left(\frac{1 + \tanh(\frac{W_{atm}}{5})}{2} \right)^{8.8} \quad (2.479)$$

where $\Delta \rho_{sno}$ (kg m^{-3}) is the increase in snow density relative to (2.478).

The mass of snow W_{sno} is

$$W_{sno}^{n+1} = W_{sno}^n + q_{grnd, ice} \Delta t. \quad (2.480)$$

The ice content of the top layer and the layer thickness are updated as

$$w_{ice, snl+1}^{n+1} = w_{ice, snl+1}^n + q_{grnd, ice} \Delta t \quad (2.481)$$

$$\Delta z_{snl+1}^{n+1} = \Delta z_{snl+1}^n + \Delta z_{sno}. \quad (2.482)$$

Since wetlands are modeled as columns of water (no soil), snow is not allowed to accumulate if the surface temperature is above freezing ($T_g > T_f$). In this case, the incoming solid precipitation is assigned to the runoff term q_{rgwl} (section 2.7.6).

In the second step, after surface fluxes and snow/soil temperatures have been determined (Chapters 2.5 and 2.6), $w_{ice, snl+1}$ is updated for frost or sublimation as

$$w_{ice, snl+1}^{n+1} = w_{ice, snl+1}^n + f_{sno} (q_{frost} - q_{subl}) \Delta t. \quad (2.483)$$

If $w_{ice, snl+1}^{n+1} < 0$ upon solution of equation , the ice content is reset to zero and the liquid water content $w_{liq, snl+1}$ is reduced by the amount required to bring $w_{ice, snl+1}^{n+1}$ up to zero.

The snow water equivalent W_{sno} is capped to not exceed 1000 kg m⁻². If the addition of q_{frost} were to result in $W_{sno} > 1000$ kg m⁻², the frost term q_{frost} is instead added to the ice runoff term $q_{snwcp, ice}$ (section 2.7.6).

2.8.3 Water Content

The conservation equation for mass of water in snow layers is

$$\frac{\partial w_{liq, i}}{\partial t} = (q_{liq, i-1} - q_{liq, i}) + \frac{(\Delta w_{liq, i})_p}{\Delta t} \quad (2.484)$$

where $q_{liq, i-1}$ is the flow of liquid water into layer i from the layer above, $q_{liq, i}$ is the flow of water out of layer i to the layer below, $(\Delta w_{liq, i})_p / \Delta t$ is the change in liquid water due to phase change (melting rate) (section 2.6.2). For the top snow layer only,

$$q_{liq, i-1} = f_{sno} (q_{grnd, liq} + (q_{sdew} - q_{seva})) \quad (2.485)$$

where $q_{grnd, liq}$ is the rate of liquid precipitation reaching the snow (section 2.7.1), q_{seva} is the evaporation of liquid water and q_{sdew} is the liquid dew (section 2.5.4). After surface fluxes and snow/soil temperatures have been determined (Chapters 2.5 and 2.6), $w_{liq, snl+1}$ is updated for the liquid precipitation reaching the ground and dew or evaporation as

$$w_{liq, snl+1}^{n+1} = w_{liq, snl+1}^n + f_{sno} (q_{grnd, liq} + q_{sdew} - q_{seva}) \Delta t. \quad (2.486)$$

When the liquid water within a snow layer exceeds the layer's holding capacity, the excess water is added to the underlying layer, limited by the effective porosity ($1 - \theta_{ice}$) of the layer. The flow of water is assumed to be zero ($q_{liq, i} = 0$) if the effective porosity of either of the two layers ($1 - \theta_{ice, i}$ and $1 - \theta_{ice, i+1}$) is less than $\theta_{imp} = 0.05$, the water impermeable volumetric water content. Thus, water flow between layers, $q_{liq, i}$, for layers $i = snl + 1, \dots, 0$, is initially calculated as

$$q_{liq, i} = \frac{\rho_{liq} [\theta_{liq, i} - S_r (1 - \theta_{ice, i})] f_{sno} \Delta z_i}{\Delta t} \geq 0 \quad (2.487)$$

where the volumetric liquid water $\theta_{liq, i}$ and ice $\theta_{ice, i}$ contents are

$$\theta_{ice, i} = \frac{w_{ice, i}}{f_{sno} \Delta z_i \rho_{ice}} \leq 1 \quad (2.488)$$

$$\theta_{liq, i} = \frac{w_{liq, i}}{f_{sno} \Delta z_i \rho_{liq}} \leq 1 - \theta_{ice, i}, \quad (2.489)$$

and $S_r = 0.033$ is the irreducible water saturation (snow holds a certain amount of liquid water due to capillary retention after drainage has ceased ([Anderson \(1976\)](#))). The water holding capacity of the underlying layer limits the flow of water $q_{liq, i}$ calculated in equation , unless the underlying layer is the surface soil layer, as

$$q_{liq, i} \leq \frac{\rho_{liq} [1 - \theta_{ice, i+1} - \theta_{liq, i+1}] \Delta z_{i+1}}{\Delta t} \quad i = snl + 1, \dots, -1. \quad (2.490)$$

The liquid water content $w_{liq, i}$ is updated as

$$w_{liq, i}^{n+1} = w_{liq, i}^n + (q_{i-1} - q_i) \Delta t. \quad (2.491)$$

Equations - are solved sequentially from top ($i = snl + 1$) to bottom ($i = 0$) snow layer in each time step. The total flow of liquid water reaching the soil surface is then $q_{liq, 0}$ which is used in the calculation of surface runoff and infiltration (sections [2.7.2](#) and [2.7.2](#)).

2.8.4 Black and organic carbon and mineral dust within snow

Particles within snow originate from atmospheric aerosol deposition (D_{sp} in Table 2.3 ($\text{kg m}^{-2} \text{ s}^{-1}$)) and influence snow radiative transfer (sections [2.3.2](#), [2.3.2](#), and [2.3.2](#)). Particle masses and mixing ratios are represented with a simple mass-conserving scheme. The model maintains masses of the following eight particle species within each snow layer: hydrophilic black carbon, hydrophobic black carbon, hydrophilic organic carbon, hydrophobic organic carbon, and four species of mineral dust with the following particle sizes: 0.1-1.0, 1.0-2.5, 2.5-5.0, and 5.0-10.0 μm . Each of these species has unique optical properties ([Table 2.12](#)) and meltwater removal efficiencies ([Table 2.19](#)).

The black carbon and organic carbon deposition rates described in Table 2.3 are combined into four categories as follows

$$D_{bc, hphil} = D_{bc, dryhphil} + D_{bc, wethphil} \quad (2.492)$$

$$D_{bc, hphob} = D_{bc, dryhphob} \quad (2.493)$$

$$D_{oc, hphil} = D_{oc, dryhphil} + D_{oc, wethphil} \quad (2.494)$$

$$D_{oc, hphob} = D_{oc, dryhphob} \quad (2.495)$$

Deposited particles are assumed to be instantly mixed (homogeneously) within the surface snow layer and are added after the inter-layer water fluxes are computed (section [2.8.3](#)) so that some aerosol is in the top layer after deposition and is not immediately washed out before radiative calculations are done. Particle masses are then redistributed each time step based on meltwater drainage through the snow column (section [2.8.3](#)) and snow layer combination and subdivision (section [2.8.7](#)). The change in mass of each of the particle species $\Delta m_{sp, i}$ (kg m^{-2}) is

$$\Delta m_{sp, i} = [k_{sp} (q_{liq, i-1} c_{sp, i-1} - q_{liq, i} c_i) + D_{sp}] \Delta t \quad (2.496)$$

where k_{sp} is the meltwater scavenging efficiency that is unique for each species (Table 2.19), $q_{liq, i-1}$ is the flow of liquid water into layer i from the layer above, $q_{liq, i}$ is the flow of water out of layer i into the layer below ($\text{kg m}^{-2} \text{s}^{-1}$) (section 2.8.3), $c_{sp, i-1}$ and $c_{sp, i}$ are the particle mass mixing ratios in layers $i - 1$ and i (kg kg^{-1}), D_{sp} is the atmospheric deposition rate (zero for all layers except layer $snl + 1$), and Δt is the model time step (s). The particle mass mixing ratio is

$$c_i = \frac{m_{sp, i}}{w_{liq, i} + w_{ice, i}}. \quad (2.497)$$

Values of k_{sp} are partially derived from experiments published by *Conway et al. (1996)*. Particles masses are re-distributed proportionately with snow mass when layers are combined or divided, thus conserving particle mass within the snow column. The mass of particles carried out with meltwater through the bottom snow layer is assumed to be permanently lost from the snowpack, and is not maintained within the model.

Table 2.19: Meltwater scavenging efficiency for particles within snow

Species	k_{sp}
Hydrophilic black carbon	0.20
Hydrophobic black carbon	0.03
Hydrophilic organic carbon	0.20
Hydrophobic organic carbon	0.03
Dust species 1 (0.1-1.0 μm)	0.02
Dust species 2 (1.0-2.5 μm)	0.02
Dust species 3 (2.5-5.0 μm)	0.01
Dust species 4 (5.0-10.0 μm)	0.01

2.8.5 Initialization of snow layer

If there are no existing snow layers ($snl + 1 = 1$) but $z_{sno} \geq 0.01 \text{ m}$ after accounting for solid precipitation q_{sno} , then a snow layer is initialized ($snl = -1$) as follows

$$\begin{aligned} \Delta z_0 &= z_{sno} \\ z_o &= -0.5\Delta z_0 \\ z_{h, -1} &= -\Delta z_0 \\ T_0 &= \min(T_f, T_{atm}) \\ w_{ice, 0} &= W_{sno} \\ w_{liq, 0} &= 0 \end{aligned} \quad (2.498)$$

2.8.6 Snow Compaction

Snow compaction is initiated after the soil hydrology calculations [surface runoff (section 2.7.2), infiltration (section 2.7.2), soil water (section 2.7.3)] are complete. Compaction of snow includes three types of processes: destructive metamorphism of new snow (crystal breakdown due to wind or thermodynamic stress); snow load or overburden (pressure); and melting (changes in snow structure due to melt-freeze cycles plus changes in crystals due to liquid water). The total fractional compaction rate for each snow layer $C_{R, i}$ (s^{-1}) is the sum of the three compaction processes

$$C_{R, i} = \frac{1}{\Delta z_i} \frac{\partial \Delta z_i}{\partial t} = C_{R1, i} + C_{R2, i} + C_{R3, i}. \quad (2.499)$$

Compaction is not allowed if the layer is saturated

$$1 - \left(\frac{w_{ice, i}}{f_{sno} \Delta z_i \rho_{ice}} + \frac{w_{liq, i}}{f_{sno} \Delta z_i \rho_{liq}} \right) \leq 0.001 \quad (2.500)$$

or if the ice content is below a minimum value ($w_{ice, i} \leq 0.1$).

Compaction as a result of destructive metamorphism $C_{R1, i}$ (s⁻¹) is temperature dependent ([Anderson \(1976\)](#))

$$C_{R1, i} = \left[\frac{1}{\Delta z_i} \frac{\partial \Delta z_i}{\partial t} \right]_{metamorphism} = -c_3 c_1 c_2 \exp [-c_4 (T_f - T_i)] \quad (2.501)$$

where $c_3 = 2.777 \times 10^{-6}$ (s⁻¹) is the fractional compaction rate for $T_i = T_f$, $c_4 = 0.04$ K⁻¹, and

$$\begin{aligned} c_1 &= 1 & \frac{w_{ice, i}}{f_{sno} \Delta z_i} &\leq 100 \text{ kg m}^{-3} \\ c_1 &= \exp \left[-0.046 \left(\frac{w_{ice, i}}{f_{sno} \Delta z_i} - 100 \right) \right] & \frac{w_{ice, i}}{f_{sno} \Delta z_i} &> 100 \text{ kg m}^{-3} \\ c_2 &= 2 & \frac{w_{liq, i}}{f_{sno} \Delta z_i} &> 0.01 \\ c_2 &= 1 & \frac{w_{liq, i}}{f_{sno} \Delta z_i} &\leq 0.01 \end{aligned} \quad (2.502)$$

where $w_{ice, i}/(f_{sno} \Delta z_i)$ and $w_{liq, i}/(f_{sno} \Delta z_i)$ are the bulk densities of liquid water and ice (kg m⁻³).

The compaction rate as a result of overburden $C_{R2, i}$ (s⁻¹) is a linear function of the snow load pressure $P_{s, i}$ (kg m⁻²) ([Anderson \(1976\)](#))

$$C_{R2, i} = \left[\frac{1}{\Delta z_i} \frac{\partial \Delta z_i}{\partial t} \right]_{overburden} = -\frac{P_{s, i}}{\eta} \quad (2.503)$$

where η is a viscosity coefficient (kg s m⁻²) that varies with density and temperature as

$$\eta = \eta_0 \exp \left[c_5 (T_f - T_i) + c_6 \frac{w_{ice, i}}{f_{sno} \Delta z_i} \right] \quad (2.504)$$

where $\eta_0 = 9 \times 10^5$ kg s m⁻², and $c_5 = 0.08$ K⁻¹, $c_6 = 0.023$ m³ kg⁻¹ are constants. The snow load pressure $P_{s, i}$ is calculated for each layer as the sum of the ice $w_{ice, i}$ and liquid water contents $w_{liq, i}$ of the layers above plus half the ice and liquid water contents of the layer being compacted

$$P_{s, i} = \frac{w_{ice, i} + w_{liq, i}}{2} + \sum_{j=snl+1}^{j=i-1} (w_{ice, j} + w_{liq, j}). \quad (2.505)$$

The compaction rate due to melting $C_{R3, i}$ (s⁻¹) is taken to be the ratio of the change in snow ice mass after the melting to the mass before melting

$$C_{R3, i} = \left[\frac{1}{\Delta z_i} \frac{\partial \Delta z_i}{\partial t} \right]_{melt} = -\frac{1}{\Delta t} \max \left(0, \frac{W_{sno, i}^n - W_{sno, i}^{n+1}}{W_{sno, i}^n} \right) \quad (2.506)$$

and melting is identified during the phase change calculations (section 2.6.2). Because snow depth is defined as the average depth of the snow covered area, the snow depth must also be updated for changes in f_{sno} .

$$C_{R4, i} = \left[\frac{1}{\Delta z_i} \frac{\partial \Delta z_i}{\partial t} \right]_{f_{sno}} = -\frac{1}{\Delta t} \max \left(0, \frac{f_{sno, i}^n - f_{sno, i}^{n+1}}{f_{sno, i}^n} \right) \quad (2.507)$$

The snow layer thickness after compaction is then

$$\Delta z_i^{n+1} = \Delta z_i^n (1 + C_{R, i} \Delta t). \quad (2.508)$$

2.8.7 Snow Layer Combination and Subdivision

After the determination of snow temperature including phase change (Chapter 2.6), snow hydrology (Chapter 2.8), and the compaction calculations (section 2.8.6), the number of snow layers is adjusted by either combining or subdividing layers. The combination and subdivision of snow layers is based on [Jordan \(1991\)](#).

Combination

If a snow layer has nearly melted or if its thickness Δz_i is less than the prescribed minimum thickness Δz_{\min} (Table 2.20), the layer is combined with a neighboring layer. The overlying or underlying layer is selected as the neighboring layer according to the following rules

1. If the top layer is being removed, it is combined with the underlying layer
2. If the underlying layer is not snow (i.e., it is the top soil layer), the layer is combined with the overlying layer
3. If the layer is nearly completely melted, the layer is combined with the underlying layer
4. If none of the above rules apply, the layer is combined with the thinnest neighboring layer.

A first pass is made through all snow layers to determine if any layer is nearly melted ($w_{ice, i} \leq 0.1$). If so, the remaining liquid water and ice content of layer i is combined with the underlying neighbor $i + 1$ as

$$w_{liq, i+1} = w_{liq, i+1} + w_{liq, i} \quad (2.509)$$

$$w_{ice, i+1} = w_{ice, i+1} + w_{ice, i}. \quad (2.510)$$

This includes the snow layer directly above the top soil layer. In this case, the liquid water and ice content of the melted snow layer is added to the contents of the top soil layer. The layer properties, T_i , $w_{ice, i}$, $w_{liq, i}$, Δz_i , are then re-indexed so that the layers above the eliminated layer are shifted down by one and the number of snow layers is decremented accordingly.

At this point, if there are no explicit snow layers remaining ($snl = 0$), the snow water equivalent W_{sno} and snow depth z_{sno} are set to zero, otherwise, W_{sno} and z_{sno} are re-calculated as

$$W_{sno} = \sum_{i=snl+1}^{i=0} (w_{ice, i} + w_{liq, i}) \quad (2.511)$$

$$z_{sno} = \sum_{i=snl+1}^{i=0} \Delta z_i. \quad (2.512)$$

If the snow depth $0 < z_{sno} < 0.01$ m or the snow density $\frac{W_{sno}}{f_{sno} z_{sno}} < 50$ kg/m³, the number of snow layers is set to zero, the total ice content of the snowpack $\sum_{i=snl+1}^{i=0} w_{ice, i}$ is assigned to W_{sno} , and the total liquid water $\sum_{i=snl+1}^{i=0} w_{liq, i}$ is assigned to the top soil layer. Otherwise, the layers are combined according to the rules above.

When two snow layers are combined (denoted here as 1 and 2), their thickness combination (c) is

$$\Delta z_c = \Delta z_1 + \Delta z_2, \quad (2.513)$$

their mass combination is

$$w_{liq, c} = w_{liq, 1} + w_{liq, 2} \quad (2.514)$$

$$w_{ice, c} = w_{ice, 1} + w_{ice, 2}, \quad (2.515)$$

and their temperatures are combined as

$$T_c = T_f + \frac{h_c - L_f w_{liq, c}}{C_{ice} w_{ice, c} + C_{liq} w_{liq, c}} \quad (2.516)$$

where $h_c = h_1 + h_2$ is the combined enthalpy h_i of the two layers where

$$h_i = (C_{ice}w_{ice,i} + C_{liq}w_{liq,i})(T_i - T_f) + L_f w_{liq,i}. \quad (2.517)$$

In these equations, L_f is the latent heat of fusion (J kg^{-1}) and C_{liq} and C_{ice} are the specific heat capacities ($\text{J kg}^{-1} \text{K}^{-1}$) of liquid water and ice, respectively (Table 2.7). After layer combination, the node depths and layer interfaces (Figure 2.9) are recalculated from

$$z_i = z_{h,i} - 0.5\Delta z_i \quad i = 0, \dots, snl + 1 \quad (2.518)$$

$$z_{h,i-1} = z_{h,i} - \Delta z_i \quad i = 0, \dots, snl + 1 \quad (2.519)$$

where Δz_i is the layer thickness.

Table 2.20: Minimum and maximum thickness of snow layers (m)

Layer	Δz_{\min}	N_l	N_u	$(\Delta z_{\max})_l$	$(\Delta z_{\max})_u$
1 (top)	0.010	1	>1	0.03	0.02
2	0.015	2	>2	0.07	0.05
3	0.025	3	>3	0.18	0.11
4	0.055	4	>4	0.41	0.23
5 (bottom)	0.115	5	•	•	•

The maximum snow layer thickness, Δz_{\max} , depends on the number of layers, N_l and N_u (section 2.8.7).

Subdivision

The snow layers are subdivided when the layer thickness exceeds the prescribed maximum thickness Δz_{\max} with lower and upper bounds that depend on the number of snow layers (numref:Table snow layer thickness). For example, if there is only one layer, then the maximum thickness of that layer is 0.03 m, however, if there is more than one layer, then the maximum thickness of the top layer is 0.02 m. Layers are checked sequentially from top to bottom for this limit. If there is only one snow layer and its thickness is greater than 0.03 m (Table 2.20), the layer is subdivided into two layers of equal thickness, liquid water and ice contents, and temperature. If there is an existing layer below the layer to be subdivided, the thickness Δz_i , liquid water and ice contents, $w_{liq,i}$ and $w_{ice,i}$, and temperature T_i of the excess snow are combined with the underlying layer according to equations -. If there is no underlying layer after adjusting the layer for the excess snow, the layer is subdivided into two layers of equal thickness, liquid water and ice contents. The vertical snow temperature profile is maintained by calculating the slope between the layer above the splitting layer (T_1) and the splitting layer (T_2) and constraining the new temperatures (T_2^{n+1}, T_3^{n+1}) to lie along this slope. The temperature of the lower layer is first evaluated from

$$T'_3 = T_2^n - \left(\frac{T_1^n - T_2^n}{(\Delta z_1^n + \Delta z_2^n)/2} \right) \left(\frac{\Delta z_2^{n+1}}{2} \right), \quad (2.520)$$

then adjusted as,

$$\begin{aligned} T_3^{n+1} &= T_2^n & T'_3 &\geq T_f \\ T_2^{n+1} &= T_2^n + \left(\frac{T_1^n - T_2^n}{(\Delta z_1^n + \Delta z_2^n)/2} \right) \left(\frac{\Delta z_2^{n+1}}{2} \right) & T'_3 &< T_f \end{aligned} \quad (2.521)$$

where here the subscripts 1, 2, and 3 denote three layers numbered from top to bottom. After layer subdivision, the node depths and layer interfaces are recalculated from equations and .

2.9 Stomatal Resistance and Photosynthesis

2.9.1 Summary of CLM5.0 updates relative to the CLM4.5

We describe here the complete photosynthesis and stomatal conductance parameterizations that appear in CLM5.0. Corresponding information for CLM4.5 appeared in the CLM4.5 Technical Note ([Oleson et al. 2013](#)).

CLM5 includes the following new changes to photosynthesis and stomatal conductance:

- Default stomatal conductance calculation uses the Medlyn conductance model
- J_{max} is predicted by the LUNA model (Chapter 2.10)
- Leaf N concentration and the fraction of leaf N in Rubisco used to calculate V_{cmax25} are determined by the LUNA model (Chapter 2.10)
- Water stress is applied by the hydraulic conductance model (Chapter 2.11)

2.9.2 Introduction

Leaf stomatal resistance, which is needed for the water vapor flux (Chapter 2.5), is coupled to leaf photosynthesis similar to Collatz et al. (1991, 1992). These equations are solved separately for sunlit and shaded leaves using average absorbed photosynthetically active radiation for sunlit and shaded leaves [ϕ^{sun}, ϕ^{sha} W m⁻² (section 2.4.1)] to give sunlit and shaded stomatal resistance (r_s^{sun}, r_s^{sha} s m⁻¹) and photosynthesis (A^{sun}, A^{sha} $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$). Canopy photosynthesis is $A^{sun}L^{sun} + A^{sha}L^{sha}$, where L^{sun} and L^{sha} are the sunlit and shaded leaf area indices (section 2.4.1). Canopy conductance is $\frac{1}{r_b+r_s^{sun}}L^{sun} + \frac{1}{r_b+r_s^{sha}}L^{sha}$, where r_b is the leaf boundary layer resistance (section 2.5.3). The implementation is described by Bonan et al. (2011), though different methods of calculating stomatal conductance, J_{max} , and the nitrogen variables used to calculate V_{cmax25} are used in CLM5.

2.9.3 Stomatal resistance

CLM5 calculates stomatal conductance using the Medlyn stomatal conductance model ([Medlyn et al. 2011](#)). Previous versions of CLM calculated leaf stomatal resistance is using the Ball-Berry conductance model as described by [Collatz et al. \(1991\)](#) and implemented in global climate models ([Sellers et al. 1996](#)). The Medlyn model calculates stomatal conductance (i.e., the inverse of resistance) based on net leaf photosynthesis, the vapor pressure deficit, and the CO₂ concentration at the leaf surface. Leaf stomatal resistance is:

$$\frac{1}{r_s} = g_s = g_o + \left(1 + \frac{g_1}{\sqrt{D}}\right) \frac{A_n}{c_s/P_{atm}} \quad (2.522)$$

where r_s is leaf stomatal resistance (s m⁻² μmol^{-1}), g_o is the minimum stomatal conductance ($\mu\text{mol m}^{-2} \text{ s}^{-1}$), A_n is leaf net photosynthesis ($\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$), c_s is the CO₂ partial pressure at the leaf surface (Pa), P_{atm} is the atmospheric pressure (Pa), and D is the vapor pressure deficit at the leaf surface (Pa). g_1 is a plant functional type dependent parameter (Table 2.21) and can be calculated as

$$g_1 = \sqrt{\frac{3\Gamma\lambda}{1.6}} \quad (2.523)$$

where Γ (mol mol⁻¹) is the CO₂ compensation point for photosynthesis without dark respiration, 1.6 is the ratio of diffusivity of CO₂ to H₂O and λ (mol H₂O mol⁻¹) is a parameter describing the marginal water cost of carbon gain. The value for $g_o = 100 \mu\text{mol m}^{-2} \text{ s}^{-1}$ for C₃ and C₄ plants. Photosynthesis is calculated for sunlit (A^{sun}) and shaded (A^{sha}) leaves to give r_s^{sun} and r_s^{sha} . Additionally, soil water influences stomatal resistance through plant hydraulic stress, detailed in the [Plant Hydraulics](#) chapter.

Resistance is converted from units of $\text{s m}^2 \mu \text{mol}^{-1}$ to s m^{-1} as: $1 \text{ s m}^{-1} = 1 \times 10^{-9} R_{\text{gas}} \frac{\theta_{\text{atm}}}{P_{\text{atm}}} \mu \text{mol}^{-1} \text{ m}^2 \text{ s}$, [same as 4.5, but units seem off. check that units are correct] where R_{gas} is the universal gas constant ($\text{J K}^{-1} \text{ kmol}^{-1}$) (Table 2.7) and θ_{atm} is the atmospheric potential temperature (K).

Table 2.21: Plant functional type (PFT) stomatal conductance parameters.

PFT	g_1 (Pa)
NET Temperate	2.35
NET Boreal	2.35
NDT Boreal	2.35
BET Tropical	4.12
BET temperate	4.12
BDT tropical	4.45
BDT temperate	4.45
BDT boreal	4.45
BES temperate	4.70
BDS temperate	4.70
BDS boreal	4.70
C ₃ arctic grass	2.22
C ₃ grass	5.25
C ₄ grass	1.62
Temperate Corn	1.79
Spring Wheat	5.79
Temperate Soybean	5.79
Cotton	5.79
Rice	5.79
Sugarcane	1.79
Tropical Corn	1.79
Tropical Soybean	5.79

2.9.4 Photosynthesis

Photosynthesis in C₃ plants is based on the model of [Farquhar et al. \(1980\)](#). Photosynthesis in C₄ plants is based on the model of [Collatz et al. \(1992\)](#). [Bonan et al. \(2011\)](#) describe the implementation, modified here. In its simplest form, leaf net photosynthesis after accounting for respiration (R_d) is

$$A_n = \min(A_c, A_j, A_p) - R_d. \quad (2.524)$$

The RuBP carboxylase (Rubisco) limited rate of carboxylation A_c ($\mu \text{ mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$) is

$$A_c = \left\{ \begin{array}{ll} \frac{V_{c \max}(c_i - \Gamma)}{c_i + K_c(1 + o_i / K_o)} & \text{for C}_3 \text{ plants} \\ \frac{V_{c \max}}{4c_i + 8\Gamma} & \text{for C}_4 \text{ plants} \end{array} \right\} \quad c_i - \Gamma \geq 0. \quad (2.525)$$

The maximum rate of carboxylation allowed by the capacity to regenerate RuBP (i.e., the light-limited rate) A_j ($\mu \text{ mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$) is

$$A_j = \left\{ \begin{array}{ll} \frac{J_x(c_i - \Gamma)}{4c_i + 8\Gamma} & \text{for C}_3 \text{ plants} \\ \alpha(4.6\phi) & \text{for C}_4 \text{ plants} \end{array} \right\} \quad c_i - \Gamma \geq 0. \quad (2.526)$$

The product-limited rate of carboxylation for C₃ plants and the PEP carboxylase-limited rate of carboxylation for C₄ plants A_p ($\mu \text{ mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$) is

$$A_p = \left\{ \begin{array}{ll} \frac{3T_p}{k_p \frac{c_i}{P_{\text{atm}}}} & \text{for C}_3 \text{ plants} \\ & \text{for C}_4 \text{ plants} \end{array} \right\}. \quad (2.527)$$

In these equations, c_i is the internal leaf CO₂ partial pressure (Pa) and $o_i = 0.20P_{atm}$ is the O₂ partial pressure (Pa). K_c and K_o are the Michaelis-Menten constants (Pa) for CO₂ and O₂. Γ (Pa) is the CO₂ compensation point. $V_{c\max}$ is the maximum rate of carboxylation ($\mu\text{mol m}^{-2} \text{s}^{-1}$, Chapter 2.10) and J_x is the electron transport rate ($\mu\text{mol m}^{-2} \text{s}^{-1}$). T_p is the triose phosphate utilization rate ($\mu\text{mol m}^{-2} \text{s}^{-1}$), taken as $T_p = 0.167V_{c\max}$ so that $A_p = 0.5V_{c\max}$ for C₃ plants (as in *Collatz et al.* 1992). For C₄ plants, the light-limited rate A_j varies with ϕ in relation to the quantum efficiency ($\alpha = 0.05 \text{ mol CO}_2 \text{ mol}^{-1} \text{ photon}$). ϕ is the absorbed photosynthetically active radiation (W m⁻²) (section 2.4.1), which is converted to photosynthetic photon flux assuming 4.6 $\mu\text{ mol photons per joule}$. k_p is the initial slope of C₄ CO₂ response curve.

For C₃ plants, the electron transport rate depends on the photosynthetically active radiation absorbed by the leaf. A common expression is the smaller of the two roots of the equation

$$\Theta_{PSII}J_x^2 - (I_{PSII} + J_{\max})J_x + I_{PSII}J_{\max} = 0 \quad (2.528)$$

where J_{\max} is the maximum potential rate of electron transport ($\mu\text{mol m}^{-2} \text{s}^{-1}$, Chapter 2.10), I_{PSII} is the light utilized in electron transport by photosystem II ($\mu\text{mol m}^{-2} \text{s}^{-1}$), and Θ_{PSII} is a curvature parameter. For a given amount of photosynthetically active radiation absorbed by a leaf (ϕ , W m⁻²), converted to photosynthetic photon flux density with 4.6 $\mu\text{ mol J}^{-1}$, the light utilized in electron transport is

$$I_{PSII} = 0.5\Phi_{PSII}(4.6\phi) \quad (2.529)$$

where Φ_{PSII} is the quantum yield of photosystem II, and the term 0.5 arises because one photon is absorbed by each of the two photosystems to move one electron. Parameter values are $\Theta_{PSII} = 0.7$ and $\Phi_{PSII} = 0.85$. In calculating A_j (for both C₃ and C₄ plants), $\phi = \phi^{sun}$ for sunlit leaves and $\phi = \phi^{sha}$ for shaded leaves.

The model uses co-limitation as described by *Collatz et al.* (1991, 1992). The actual gross photosynthesis rate, A , is given by the smaller root of the equations

$$\begin{aligned} \Theta_{cj}A_i^2 - (A_c + A_j)A_i + A_cA_j &= 0 \\ \Theta_{ip}A^2 - (A_i + A_p)A + A_iA_p &= 0 \end{aligned} \quad (2.530)$$

Values are $\Theta_{cj} = 0.98$ and $\Theta_{ip} = 0.95$ for C₃ plants; and $\Theta_{cj} = 0.80$ and $\Theta_{ip} = 0.95$ for C₄ plants. A_i is the intermediate co-limited photosynthesis. $A_n = A - R_d$.

The parameters K_c , K_o , and Γ depend on temperature. Values at 25 °C are $K_{c25} = 404.9 \times 10^{-6}P_{atm}$, $K_{o25} = 278.4 \times 10^{-3}P_{atm}$, and $\Gamma_{25} = 42.75 \times 10^{-6}P_{atm}$. $V_{c\max}$, J_{\max} , T_p , k_p , and R_d also vary with temperature. Parameter values at 25 °C are calculated from $V_{c\max}$ at 25 °C: $J_{\max 25} = 1.97V_{c\max 25}$, $T_{p25} = 0.167V_{c\max 25}$, and $R_{d25} = 0.015V_{c\max 25}$ (C₃) and $R_{d25} = 0.025V_{c\max 25}$ (C₄). For C₄ plants, $k_{p25} = 20000 V_{c\max 25}$. However, when the biogeochemistry is active, R_{d25} is calculated from leaf nitrogen as $R_{d25} = 0.2577LNC_a$, where LNC_a is the area-based leaf nitrogen concentration (g N m⁻² leaf area, Chapter 2.10), and 0.2577 $\mu\text{mol CO}_2 \text{ g}^{-1} \text{ N s}^{-1}$ is the base respiration rate. [this doesn't look correct based on the code, which lists two options; verify with Rosie] The parameters $V_{c\max 25}$, $J_{\max 25}$, T_{p25} , k_{p25} , and R_{d25} are scaled over the canopy for sunlit and shaded leaves (section 2.9.5). In C₃ plants, these are adjusted for leaf temperature, T_v (K), as:

$$\begin{aligned} V_{c\max} &= V_{c\max 25} f(T_v) f_H(T_v) \\ J_{\max} &= J_{\max 25} f(T_v) f_H(T_v) \\ T_p &= T_{p25} f(T_v) f_H(T_v) \\ R_d &= R_{d25} f(T_v) f_H(T_v) \\ K_c &= K_{c25} f(T_v) \\ K_o &= K_{o25} f(T_v) \\ \Gamma &= \Gamma_{25} f(T_v) \end{aligned} \quad (2.531)$$

with

$$f(T_v) = \exp \left[\frac{\Delta H_a}{298.15 \times 0.001R_{gas}} \left(1 - \frac{298.15}{T_v} \right) \right] \quad (2.532)$$

and

$$f_H(T_v) = \frac{1 + \exp\left(\frac{298.15\Delta S - \Delta H_d}{298.15 \times 0.001 R_{gas}}\right)}{1 + \exp\left(\frac{\Delta S T_v - \Delta H_d}{0.001 R_{gas} T_v}\right)}. \quad (2.533)$$

[Table 2.22](#) lists parameter values for ΔH_a and ΔH_d . ΔS is calculated separately for $V_{c\max}$ and J_{\max} to allow for temperature acclimation of photosynthesis (see equation (2.537)), and ΔS is $490 \text{ J mol}^{-1} \text{ K}^{-1}$ for R_d ([Bonan et al. 2011](#), [Lombardozzi et al. 2015](#)). Because T_p as implemented here varies with $V_{c\max}$, T_p uses the same temperature parameters as $V_{c\max}$. For C₄ plants,

$$\begin{aligned} V_{c\max} &= V_{c\max 25} \left[\frac{Q_{10}^{(T_v - 298.15)/10}}{f_H(T_v) f_L(T_v)} \right] \\ f_H(T_v) &= 1 + \exp[s_1(T_v - s_2)] \\ f_L(T_v) &= 1 + \exp[s_3(s_4 - T_v)] \end{aligned} \quad (2.534)$$

with $Q_{10} = 2$, $s_1 = 0.3 \text{ K}^{-1}$, $s_2 = 313.15 \text{ K}$, $s_3 = 0.2 \text{ K}^{-1}$, and $s_4 = 288.15 \text{ K}$. Additionally,

$$R_d = R_{d25} \left\{ \frac{Q_{10}^{(T_v - 298.15)/10}}{1 + \exp[s_5(T_v - s_6)]} \right\} \quad (2.535)$$

with $Q_{10} = 2$, $s_5 = 1.3 \text{ K}^{-1}$ and $s_6 = 328.15 \text{ K}$, and

$$k_p = k_{p25} Q_{10}^{(T_v - 298.15)/10} \quad (2.536)$$

with $Q_{10} = 2$.

[Table 2.22](#): Temperature dependence parameters for C3 photosynthesis.

Parameter	$\Delta H_a (\text{J mol}^{-1})$	$\Delta H_d (\text{J mol}^{-1})$
$V_{c\max}$	72000	200000
J_{\max}	50000	200000
T_p	72000	200000
R_d	46390	150650
K_c	79430	–
K_o	36380	–
Γ	37830	–

In the model, acclimation is implemented as in [Kattge and Knorr \(2007\)](#). In this parameterization, $V_{c\max}$ and J_{\max} vary with the plant growth temperature. This is achieved by allowing ΔS to vary with growth temperature according to

$$\begin{aligned} \Delta S &= 668.39 - 1.07(T_{10} - T_f) && \text{for } V_{c\max} \\ \Delta S &= 659.70 - 0.75(T_{10} - T_f) && \text{for } J_{\max} \end{aligned} \quad (2.537)$$

The effect is to cause the temperature optimum of $V_{c\max}$ and J_{\max} to increase with warmer temperatures. Additionally, the ratio $J_{\max 25}/V_{c\max 25}$ at 25°C decreases with growth temperature as

$$J_{\max 25}/V_{c\max 25} = 2.59 - 0.035(T_{10} - T_f). \quad (2.538)$$

In these acclimation functions, T_{10} is the 10-day mean air temperature (K) and T_f is the freezing point of water (K). For lack of data, T_p acclimates similar to $V_{c\max}$. Acclimation is restricted over the temperature range $T_{10} - T_f \geq 11^\circ\text{C}$ and $T_{10} - T_f \leq 35^\circ\text{C}$.

2.9.5 Canopy scaling

$V_{c \max 25}$ is calculated separately for sunlit and shaded leaves using an exponential profile to area-based leaf nitrogen (LNC_a , see Chapter 2.10), as in [Bonan et al. \(2011\)](#). $V_{c \max 25}$ at cumulative leaf area index x from the canopy top scales directly with LNC_a , which decreases exponentially with greater cumulative leaf area, so that [Verify with Rosie- different based on Vcmax option 0, 3, & 4]

$$V_{c \max 25}(x) = V_{c \max 25}(0) e^{-K_n x} \quad (2.539)$$

where $V_{c \max 25}(0)$ is defined at the top of the canopy using SLA_0 , which is the specific leaf area at the canopy top and K_n is the decay coefficient for nitrogen. The canopy integrated value for sunlit and shaded leaves is

$$\begin{aligned} V_{c \max 25}^{sun} &= \int_0^L V_{c \max 25}(x) f_{sun}(x) dx \\ &= V_{c \max 25}(0) \left[1 - e^{-(K_n + K)L} \right] \frac{1}{K_n + K} \end{aligned} \quad (2.540)$$

$$\begin{aligned} V_{c \max 25}^{sha} &= \int_0^L V_{c \max 25}(x) [1 - f_{sun}(x)] dx \\ &= V_{c \max 25}(0) \left\{ \left[1 - e^{-K_n L} \right] \frac{1}{K_n} - \left[1 - e^{-(K_n + K)L} \right] \frac{1}{K_n + K} \right\} \end{aligned} \quad (2.541)$$

and the average value for the sunlit and shaded leaves is

$$\bar{V}_{c \max 25}^{sun} = V_{c \max 25}^{sun} / L^{sun} \quad (2.542)$$

$$\bar{V}_{c \max 25}^{sha} = V_{c \max 25}^{sha} / L^{sha}. \quad (2.543)$$

This integration is over all leaf area (L) with $f_{sun}(x) = \exp(-Kx)$ and K the direct beam extinction coefficient (equation (2.97) in chapter 2.4). Photosynthetic parameters $J_{\max 25}$, T_{p25} , k_{p25} , and R_{d25} scale similarly.

The model uses $K_n = 0.30$ to match an explicit multi-layer canopy, as in [Bonan et al. \(2012\)](#). The value $K_n = 0.11$ chosen by [Bonan et al. \(2011\)](#) is consistent with observationally-derived estimates for forests, mostly tropical, and provides a gradient in $V_{c \max}$ similar to the original CLM4 specific leaf area scaling. However, [Bonan et al. \(2012\)](#) showed that the sunlit/shaded canopy parameterization does not match an explicit multi-layer canopy parameterization. The discrepancy arises from absorption of scattered radiation by shaded leaves and can be tuned out with higher K_n .

2.9.6 Numerical implementation

The CO₂ partial pressure at the leaf surface, c_s (Pa), and the vapor pressure at the leaf surface, e_s (Pa), needed for the stomatal resistance model in equation (2.522), and the internal leaf CO₂ partial pressure c_i (Pa), needed for the photosynthesis model in equations (2.524)-(2.526), are calculated assuming there is negligible capacity to store CO₂ and water vapor at the leaf surface so that

$$A_n = \frac{c_a - c_i}{(1.4r_b + 1.6r_s) P_{atm}} = \frac{c_a - c_s}{1.4r_b P_{atm}} = \frac{c_s - c_i}{1.6r_s P_{atm}} \quad (2.544)$$

and the transpiration fluxes are related as

$$\frac{e_a - e_i}{r_b + r_s} = \frac{e_a - e_s}{r_b} = \frac{e_s - e_i}{r_s} \quad (2.545)$$

where r_b is leaf boundary layer resistance (s m² μ mol⁻¹) (section 2.5.3), the terms 1.4 and 1.6 are the ratios of diffusivity of CO₂ to H₂O for the leaf boundary layer resistance and stomatal resistance, c_a = CO₂ (mol mol⁻¹), P_{atm} is the atmospheric CO₂ partial pressure (Pa) calculated from CO₂ concentration (ppmv), e_i is the saturation

vapor pressure (Pa) evaluated at the leaf temperature T_v , and e_a is the vapor pressure of air (Pa). The vapor pressure of air in the plant canopy e_a (Pa) is determined from

$$e_a = \frac{P_{atm}q_s}{0.622} \quad (2.546)$$

where q_s is the specific humidity of canopy air (kg kg^{-1} , section 2.5.3). Equations and are solved for c_s and e_s

$$c_s = c_a - 1.4r_b P_{atm} A_n \quad (2.547)$$

$$e_s = \frac{e_a r_s + e_i r_b}{r_b + r_s} \quad (2.548)$$

Substitution of equation (2.548) into equation (2.522) gives an expression for stomatal resistance (r_s) as a function of photosynthesis (A_n), given here in terms of conductance with $g_s = 1/r_s$ and $g_b = 1/r_b$

$$c_s g_s^2 + [c_s (g_b - b) - mA_n P_{atm}] g_s - g_b [c_s b + mA_n P_{atm} e_a / e(T_v)] = 0. \quad (2.549)$$

Stomatal conductance is the larger of the two roots that satisfy the quadratic equation. Values for c_i are given by

$$c_i = c_a - (1.4r_b + 1.6r_s) P_{atm} A_n \quad (2.550)$$

The equations for c_i , c_s , r_s , and A_n are solved iteratively until c_i converges. [Sun et al. \(2012\)](#) pointed out that the CLM4 numerical approach does not always converge. Therefore, the model uses a hybrid algorithm that combines the secant method and Brent's method to solve for c_i . The equation set is solved separately for sunlit (A_n^{sun} , r_s^{sun}) and shaded (A_n^{sha} , r_s^{sha}) leaves.

The model has an optional (though not supported) multi-layer canopy, as described by [Bonan et al. \(2012\)](#). The multi-layer model is only intended to address the non-linearity of light profiles, photosynthesis, and stomatal conductance in the plant canopy. In the multi-layer canopy, sunlit (A_n^{sun} , r_s^{sun}) and shaded (A_n^{sha} , r_s^{sha}) leaves are explicitly resolved at depths in the canopy using a light profile (Chapter 2.4). In this case, $V_{c\max 25}$ is not integrated over the canopy, but is instead given explicitly for each canopy layer using equation (2.539). This also uses the [Lloyd et al. \(2010\)](#) relationship whereby K_n scales with $V_{c\max}$ as

$$K_n = \exp(0.00963V_{c\max} - 2.43) \quad (2.551)$$

such that higher values of $V_{c\max}$ imply steeper declines in photosynthetic capacity through the canopy with respect to cumulative leaf area.

2.10 Photosynthetic Capacity

The photosynthetic capacity is represented by two key parameters: 1) the maximum rate of carboxylation at 25 °C, $V_{c,\max 25}$; and 2) the maximum rate of electron transport at 25 °C, $J_{\max 25}$. They are predicted by a mechanistic model of leaf utilization of nitrogen for assimilation (LUNA V1.0) (Ali et al. 2016) based on an optimality hypothesis to nitrogen allocation among light capture, electron transport, carboxylation, respiration and storage. Specifically, the model allocates the nitrogen by maximizing the daily net photosynthetic carbon gain under following two key assumptions:

- nitrogen allocated for light capture, electron transport and carboxylation are co-limiting;
- respiratory nitrogen is allocated to maintain dark respiration determined by $V_{c,\max}$.

Compared to traditional photosynthetic capacity models, a key advantage of LUNA is that the model is able to predict the potential acclimation of photosynthetic capacities at different environmental conditions as determined by temperature, radiation, CO₂ concentrations, day length, and humidity.

2.10.1 Model inputs and parameter estimations

The LUNA model includes the following four unitless parameters:

- J_{maxb0} , which specifies the baseline proportion of nitrogen allocated for electron transport;
- J_{maxb1} , which determines response of electron transport rate to light availability;
- $t_{c,j0}$, which defines the baseline ratio of Rubisco-limited rate to light-limited rate;
- H , which determines the response of electron transport rate to relative humidity.

The above four parameters are estimated by fitting the LUNA model to a global compilation of >800 observations located at different biomes, canopy locations, and time of the year from 1993-2013 (Ali et al 2015). The model inputs are area-based leaf nitrogen content, leaf mass per unit leaf area and the driving environmental conditions (average of past 10 days) including temperature, CO₂ concentrations, daily mean and maximum radiation, relative humidity and day length. The estimated values in CLM5 for the listed parameters are 0.0311, 0.1745, 0.8054, and 6.0999, respectively. In LUNA V1.0, the estimated parameter values are for C3 natural vegetations. In view that potentially large differences in photosynthetic capacity could exist between crops and natural vegetations due to human selection and genetic modifications, in CLM5, the LUNA model are used only for C3 natural vegetations. The photosynthetic capacity for crops and C4 plants are thus still kept the same as CLM4.5. Namely, it is estimated based on the leaf nitrogen content, fixed RUBISCO allocations for $V_{c\max}$ and an adjusting factor to account for the impact of day length. In CLM5, the model simulates both sun-lit and shaded leaves; however, because the sun-lit and shaded leaves can change through the day based on the sun angles, we do not differentiate the photosynthetic capacity difference for sun-lit or shaded leaves.

2.10.2 Model structure

Plant Nitrogen

The structure of the LUNA model is adapted from Xu et al.(2012), where the plant nitrogen at the leaf level (LNC_a; gN/m² leaf) is divided into four pools: structural nitrogen(N_{str} ; gN/m² leaf), photosynthetic nitrogen (N_{psn} ; gN/m² leaf), storage nitrogen(N_{store} ; gN/m² leaf), and respiratory nitrogen (N_{resp} ; gN/m² leaf). Namely,

$$LNC_a = N_{psn} + N_{str} + N_{store} + N_{resp}. \quad (2.552)$$

The photosynthetic nitrogen, N_{psn} , is further divided into nitrogen for light capture (N_{lc} ; gN/m² leaf), nitrogen for electron transport (N_{et} ; gN/m² leaf), and nitrogen for carboxylation (N_{cb} ; gN/m² leaf). Namely,

$$N_{psn} = N_{et} + N_{cb} + N_{lc}. \quad (2.553)$$

The structural nitrogen, N_{str} , is calculated as the multiplication of leaf mass per unit area (LMA: g biomass/m² leaf), and the structural nitrogen content (SNC; gN/g biomass). Namely,

$$N_{str} = SNC \cdot LMA \quad (2.554)$$

where SNC is set to be fixed at 0.002 (gN/g biomass), based on data on C:N ratio from dead wood (White et al.,2000).

We assume that plants optimize their nitrogen allocations (i.e., N_{store} , N_{resp} , N_{lc} , N_{et} , N_{cb}) to maximize the photosynthetic carbon gain, defined as the gross photosynthesis (A) minus the maintenance respiration for photosynthetic enzymes (R_{psn}), under specific environmental conditions and given plant's strategy of leaf nitrogen use. Namely, the solutions of nitrogen allocations { N_{store} , N_{resp} , N_{lc} , N_{et} , N_{cb} } can be estimated as follows,

$$\left\{ \hat{N}_{store}, \hat{N}_{resp}, \hat{N}_{lc}, \hat{N}_{et}, \hat{N}_{cb} \right\} = \underset{N_{store} + N_{resp} + N_{lc} + N_{et} + N_{cb} < FNC_a}{\operatorname{argmax}} (A - R_{psn}), \quad (2.555)$$

where FNC_a is the functional nitrogen content defined as the total leaf nitrogen content (LNC_a) minus the structural nitrogen content (N_{str}).

The gross photosynthesis, A , was calculated with a coupled leaf gas exchange model based on the Farquhar et al. (1980) model of photosynthesis and Ball–Berry-type stomatal conductance model (Ball et al., 1987). The maintenance respiration for photosynthetic enzymes, R_{psn} , is calculated by the multiplication of total photosynthetic nitrogen (N_{psn}) and the maintenance respiration cost for photosynthetic enzymes.

Maximum electron transport rate

In the LUNA model, the maximum electron transport rate (J_{max} ; $\mu\text{mol electron} / \text{m}^{-2}/\text{s}$) is simulated to have a baseline allocation of nitrogen and additional nitrogen allocation to change depending on the average daytime photosynthetic active radiation (PAR; $\mu\text{mol electron} / \text{m}^{-2}/\text{s}$), day length (hours) and air humidity. Specifically, the LUNA model has

$$J_{\text{max}} = J_{\text{max}0} + J_{\text{max}b1} f(\text{day length}) f(\text{humidity}) \alpha \text{PAR} \quad (2.556)$$

The baseline electron transport rate, $J_{\text{max}0}$, is calculated as follows,

$$J_{\text{max}0} = J_{\text{max}b0} \text{FNC}_a \text{NUE}_{J_{\text{max}}} \quad (2.557)$$

where $J_{\text{max}b0}$ (unitless) is the baseline proportion of nitrogen allocated for electron transport rate. $\text{NUE}_{J_{\text{max}}} (\mu\text{mol electron} / \text{s/g N})$ is the nitrogen use efficiency of J_{max} . $J_{\text{max}b1}$ (unitless) is a coefficient determining the response of the electron transport rate to amount of absorbed light (i.e., αPAR). $f(\text{day length})$ is a function specifies the impact of day length (hours) on J_{max} in view that longer day length has been demonstrated by previous studies to alter $V_{\text{cmax}25}$ and $J_{\text{max}25}$ (Bauerle et al., 2012; Comstock and Ehleringer, 1986) through photoperiod sensing and regulation (e.g. Song et al., 2013). Following Bauerle et al. (2012), $f(\text{day length})$ is simulated as follows,

$$f(\text{day length}) = \left(\frac{\text{day length}}{12} \right)^2. \quad (2.558)$$

$f(\text{humidity})$ represents the impact of air humidity on J_{max} . We assume that higher humidity leads to higher J_{max} with less water limitation on stomata opening and that low relative humidity has a stronger impact on nitrogen allocation due to greater water limitation. When relative humidity (RH; unitless) is too low, we assume that plants are physiologically unable to reallocate nitrogen. We therefore assume that there exists a critical value of relative humidity ($RH_0 = 0.25$; unitless), below which there is no optimal nitrogen allocation. Based on the above assumptions, we have

$$f(\text{humidity}) = \left(1 - e^{\left(-H \frac{\max(RH - RH_0, 0)}{1 - RH_0} \right)} \right), \quad (2.559)$$

where H (unitless) specifies the impact of relative humidity on electron transport rate.

The efficiency of light energy absorption (unitless), α , is calculated depending on the amount of nitrogen allocated for light capture, N_{lc} . Following Niinemets and Tenhunen (1997), the LUNA model has,

$$\alpha = \frac{0.292}{1 + \frac{0.076}{N_{\text{lc}} C_b}} \quad (2.560)$$

where 0.292 is the conversion factor from photon to electron. C_b is the conversion factor (1.78) from nitrogen to chlorophyll. After we estimate J_{max} , the actual electron transport rate with the daily maximum radiation (J_x) can be calculated using the empirical expression of leaf (1937),

$$J_x = \frac{\alpha \text{PAR}_{\text{max}}}{\left(1 + \frac{\alpha^2 \text{PAR}_{\text{max}}^2}{J_{\text{max}}^2} \right)^{0.5}} \quad (2.561)$$

where $\text{PAR}_{\text{max}} (\mu\text{mol/m}^2/\text{s})$ is the maximum photosynthetically active radiation during the day.

Maximum rate of carboxylation

The maximum rate of carboxylation at 25 °C varies with foliage nitrogen concentration and specific leaf area and is calculated as in Thornton and Zimmermann (2007). At 25°C,

$$V_{c\max 25} = N_{str} N_{cb} F_{NR} a_{R25} \quad (2.562)$$

where N_{str} is the area-based leaf nitrogen concentration (g N m^{-2} leaf area), N_{cb} is the fraction of leaf nitrogen in Rubisco ($\text{g N in Rubisco g}^{-1}$ N), $F_{NR} = 7.16$ is the mass ratio of total Rubisco molecular mass to nitrogen in Rubisco (g Rubisco g^{-1} N in Rubisco), and $a_{R25} = 60$ is the specific activity of Rubisco ($\mu\text{mol CO}_2 \text{ g}^{-1}$ Rubisco s^{-1}).

$V_{c\max 25}$ additionally varies with daylength (DYL) using the function $f(DYL)$, which introduces seasonal variation to $V_{c\max}$

$$f(DYL) = \frac{(DYL)^2}{(DYL_{\max})^2} \quad (2.563)$$

with $0.01 \leq f(DYL) \leq 1$. Daylength (seconds) is given by

$$DYL = 2 \times 13750.9871 \cos^{-1} \left[\frac{-\sin(lat) \sin(decl)}{\cos(lat) \cos(decl)} \right] \quad (2.564)$$

where lat (latitude) and $decl$ (declination angle) are from section 2.3.3. Maximum daylength (DYL_{\max}) is calculated similarly but using the maximum declination angle for present-day orbital geometry ($\pm 23.4667^\circ$ [± 0.409571 radians], positive for Northern Hemisphere latitudes and negative for Southern Hemisphere).

Implementation of Photosynthetic Capacity

Based on Farquhar et al. (1980) and Wullschleger (1993), we can calculate the electron-limited photosynthetic rate under daily maximum radiation (W_{J_x}) and the Rubisco-limited photosynthetic rate (W_c) as follows,

$$W_{J_x} = K_j J_x, \quad (2.565)$$

$$W_c = K_c V_{c,\max}, \quad (2.566)$$

where K_j and K_c as the conversion factors for J_x and $V_{c,\max}$ ($V_{c,\max}$ to W_c and J_x to W_{J_x}), respectively. Based on Xu et al. (2012), Maire et al. (2012) and Walker et al. (2014), we assume that W_c is proportional to W_{J_x} . Specifically, we have

$$W_c = t_\alpha t_{c,j0} W_{J_x} \quad (2.567)$$

where $t_{c,j0}$ is the baseline ratio of W_c to W_{J_x} . We recognize that this ratio may change depending on the nitrogen use efficiency of carboxylation and electron transport (Ainsworth and Rogers, 2007), therefore the LUNA model has the modification factor, t_α , to adjust baseline the ratio depending on the nitrogen use efficiency for electron vs carboxylation (Ali et al 2016).

Total Respiration

Following Collatz et al.(1991a), the total respiration (R_t) is calculated in proportion to $V_{c,\max}$,

$$R_t = 0.015 V_{c,\max}. \quad (2.568)$$

Accounting for the daytime and nighttime temperature, the daily respirations is calculated as follows,

$$R_{td} = R_t [D_{\text{day}} + D_{\text{night}} f_r(T_{\text{night}})/f_r(T_{\text{day}})], \quad (2.569)$$

where D_{day} and D_{night} are daytime and nighttime durations in seconds. $f_r(T_{\text{night}})$ and $f_r(T_{\text{day}})$ are the temperature response functions for respiration [see Appendix B in Ali et al (2016) for details].

2.10.3 Numerical scheme

The LUNA model searches for the “optimal” nitrogen allocations for maximum net photosynthetic carbon gain by incrementally increase the nitrogen allocated for light capture (i.e., N_{lc}) (see Ali et al 2016 for details) . We assume that plants only optimize the nitrogen allocation when they can grow (i.e., GPP>0.0). If GPP become zero under stress, then the LUNA model assume a certain amount of enzyme will decay at daily rates of 0.1, in view that the half-life time for photosynthetic enzymes are short (~7 days) (Suzuki et al. 2001). To avoid unrealistic low values of photosynthetic capacity, the decay is only limited to 50 percent of the original enzyme levels.

2.11 Plant Hydraulics

2.11.1 Roots

Vertical Root Distribution

The root fraction r_i in each soil layer depends on the plant functional type

$$r_i = (\beta^{z_{h,i-1} \cdot 100} - \beta^{z_{h,i} \cdot 100}) \quad \text{for } 1 \leq i \leq N_{levsoi} \quad (2.570)$$

where $z_{h,i}$ (m) is the depth from the soil surface to the interface between layers i and $i + 1$ ($z_{h,0}$, the soil surface) (section 2.2.2), the factor of 100 converts from m to cm, and β is a plant-dependent root distribution parameter adopted from *Jackson et al. (1996)* (Table 2.23).

Table 2.23: Plant functional type root distribution parameters

Plant Functional Type	β
NET Temperate	0.976
NET Boreal	0.943
NDT Boreal	0.943
BET Tropical	0.993
BET temperate	0.966
BDT tropical	0.993
BDT temperate	0.966
BDT boreal	0.943
BES temperate	0.964
BDS temperate	0.964
BDS boreal	0.914
C ₃ grass arctic	0.914
C ₃ grass	0.943
C ₄ grass	0.943
Crop R	0.943
Crop I	0.943
Corn R	0.943
Corn I	0.943
Temp Cereal R	0.943
Temp Cereal I	0.943
Winter Cereal R	0.943
Winter Cereal I	0.943
Soybean R	0.943
Soybean I	0.943

Root Spacing

To determine the conductance along the soil to root pathway (section 2.11.2) an estimate of the spacing between the roots within a soil layer is required. The distance between roots $dx_{root,i}$ (m) is calculated by assuming that roots are distributed uniformly throughout the soil (*Gardner 1960*)

$$dx_{root,i} = (\pi \cdot L_i)^{-\frac{1}{2}} \quad (2.571)$$

where L_i is the root length density (m m^{-3})

$$L_i = \frac{B_{root,i}}{\rho_{root} CA_{root}}, \quad (2.572)$$

$B_{root,i}$ is the root biomass density (kg m^{-3})

$$B_{root,i} = \frac{c_to_b \cdot C_{fineroot} \cdot r_i}{dz_i} \quad (2.573)$$

where $c_to_b = 2$ ($\text{kg biomass kg carbon}^{-1}$) and $C_{fineroot}$ is the amount of fine root carbon (kg m^{-2}).

ρ_{root} is the root density (kg m^{-3}), and CA_{root} is the fine root cross sectional area (m^2)

$$CA_{root} = \pi r_{root}^2 \quad (2.574)$$

where r_{root} is the root radius (m).

2.11.2 Plant Hydraulic Stress

The Plant Hydraulic Stress (PHS) routine explicitly models water transport through the vegetation according to a simple hydraulic framework following Darcy's Law for porous media flow equations influenced by *Bonan et al. (2014)*, *Chuang et al. (2006)*, *Sperry et al. (1998)*, *Sperry and Love (2015)*, *Williams et al (1996)*.

PHS solves for the vegetation water potential that matches water supply with transpiration demand. Water supply is modeled according to the circuit analog in Figure 2.10. Transpiration demand is modeled relative to maximum transpiration by a transpiration loss function dependent on leaf water potential.

Plant Water Supply

The supply equations are used to solve for vegetation water potential forced by transpiration demand and the set of layer-by-layer soil water potentials. The water supply is discretized into segments: soil-to-root, root-to-stem, and stem-to-leaf. There are typically several (1-49) soil-to-root flows operating in parallel, one per soil layer. There are two stem-to-leaf flows operating in parallel, corresponding to the sunlit and shaded "leaves".

In general the water fluxes (e.g. soil-to-root, root-to-stem, etc.) are modeled according to Darcy's Law for porous media flow as:

$$q = kA(\psi_1 - \psi_2) \quad (2.575)$$

q is the flux of water (mmH_2O) spanning the segment between ψ_1 and ψ_2

k is the hydraulic conductance (s^{-1})

A is the area basis (m^2/m^2) relating the conducting area basis to ground area

$\psi_1 - \psi_2$ is the gradient in water potential (mmH_2O) across the segment

The segments in Figure 2.10 have variable resistance, as the water potentials become lower, hydraulic conductance decreases. This is captured by multiplying the maximum segment conductance by a sigmoidal function capturing

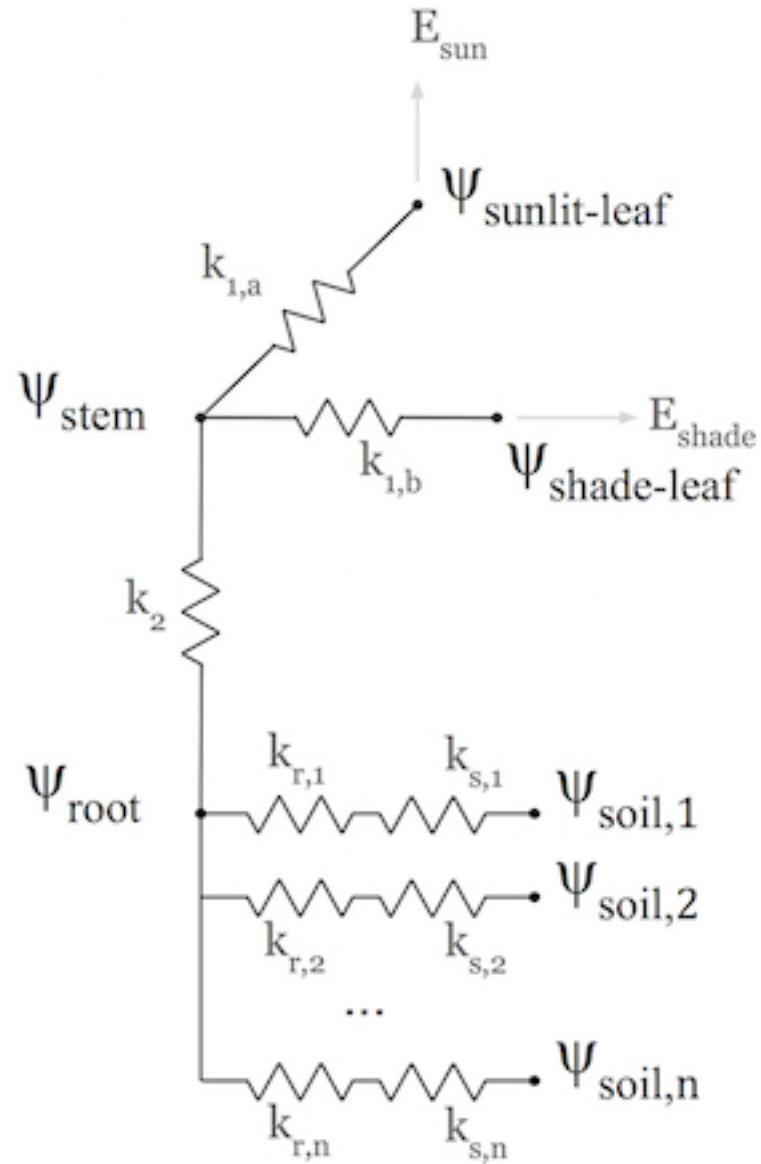


Figure 2.10: Circuit diagram of plant hydraulics scheme

the percent loss of conductivity. The function uses two parameters to fit experimental vulnerability curves: the water potential at 50% loss of conductivity ($p50$) and a shape fitting parameter (c_k).

$$k = k_{max} \cdot 2^{-\left(\frac{\psi_1}{p50}\right)^{c_k}} \quad (2.576)$$

k_{max} is the maximum segment conductance (s-1)

$p50$ is the water potential at 50% loss of conductivity (mmH₂O)

ψ_1 is the water potential of the lower segment terminus (mmH₂O)

Stem-to-leaf

The area basis and conductance parameterization varies by segment. There are two stem-to-leaf fluxes in parallel, from stem to sunlit leaf and from stem to shaded leaf (q_{1a} and q_{1b}). The water flux from stem-to-leaf is the product of the segment conductance, the conducting area basis, and the water potential gradient from stem to leaf. Stem-to-leaf conductance is defined as the maximum conductance multiplied by the percent of maximum conductance, as calculated by the sigmoidal vulnerability curve. The maximum conductance is a PFT parameter representing the maximum conductance of water from stem to leaf per unit leaf area. This parameter can be defined separately for sunlit and shaded segments and should already include the appropriate length scaling (in other words this is a conductance, not conductivity). The water potential gradient is the difference between leaf water potential and stem water potential. There is no gravity term, assuming a negligible difference in height across the segment. The area basis is the leaf area index (either sunlit or shaded).

$$q_{1a} = k_{1a} \cdot LAI_{sun} \cdot (\psi_{stem} - \psi_{sunleaf}) \quad (2.577)$$

$$q_{1b} = k_{1b} \cdot LAI_{shade} \cdot (\psi_{stem} - \psi_{shadefleaf}) \quad (2.578)$$

$$k_{1a} = k_{1a,max} \cdot 2^{-\left(\frac{\psi_{stem}}{p50_1}\right)^{c_k}} \quad (2.579)$$

$$k_{1b} = k_{1b,max} \cdot 2^{-\left(\frac{\psi_{stem}}{p50_1}\right)^{c_k}} \quad (2.580)$$

Variables:

q_{1a} = flux of water (mmH₂O/s) from stem to sunlit leaf

q_{1b} = flux of water (mmH₂O/s) from stem to shaded leaf

LAI_{sun} = sunlit leaf area index (m²/m²)

LAI_{shade} = shaded leaf area index (m²/m²)

ψ_{stem} = stem water potential (mmH₂O)

$\psi_{sunleaf}$ = sunlit leaf water potential (mmH₂O)

$\psi_{shadefleaf}$ = shaded leaf water potential (mmH₂O)

Parameters:

$k_{1a,max}$ = maximum leaf conductance (s-1)

$k_{1b,max}$ = maximum leaf conductance (s-1)

$p50_1$ = water potential at 50% loss of conductance (mmH₂O)

c_k = vulnerability curve shape-fitting parameter (-)

Root-to-stem

There is one root-to-stem flux. This represents a flux from the root collar to the upper branch reaches. The water flux from root-to-stem is the product of the segment conductance, the conducting area basis, and the water potential gradient from root to stem. Root-to-stem conductance is defined as the maximum conductance multiplied by the percent of maximum conductance, as calculated by the sigmoidal vulnerability curve (two parameters). The maximum conductance is defined as the maximum root-to-stem conductivity per unit stem area (PFT parameter) divided by the length of the conducting path, which is taken to be the vegetation height. The area basis is the stem area index. The gradient in water potential is the difference between the root water potential and the stem water potential less the difference in gravitational potential.

$$q_2 = k_2 \cdot SAI \cdot (\psi_{root} - \psi_{stem} - \Delta\psi_z) \quad (2.581)$$

$$k_2 = \frac{k_{2,max}}{z_2} \cdot 2^{-\left(\frac{\psi_{root}}{p50_2}\right)^{c_k}} \quad (2.582)$$

Variables:

q_2 = flux of water (mmH₂O/s) from root to stem

SAI = stem area index (m²/m²)

$\Delta\psi_z$ = gravitational potential (mmH₂O)

ψ_{root} = root water potential (mmH₂O)

ψ_{stem} = stem water potential (mmH₂O)

Parameters:

$k_{2,max}$ = maximum stem conductivity (m/s)

$p50_2$ = water potential at 50% loss of conductivity (mmH₂O)

z_2 = vegetation height (m)

Soil-to-root

There are several soil-to-root fluxes operating in parallel (one for each root-containing soil layer). Each represents a flux from the given soil layer to the root collar. The water flux from soil-to-root is the product of the segment conductance, the conducting area basis, and the water potential gradient from soil to root. The area basis is a proxy for root area index, defined as the summed leaf and stem area index multiplied by the root-to-shoot ratio (PFT parameter) multiplied by the layer root fraction. The root fraction comes from an empirical root profile (section 2.11.1).

The gradient in water potential is the difference between the soil water potential and the root water potential less the difference in gravitational potential. There is only one root water potential to which all soil layers are connected in parallel. A soil-to-root flux can be either positive (vegetation water uptake) or negative (water deposition), depending on the relative values of the root and soil water potentials. This allows for the occurrence of hydraulic redistribution where water moves through vegetation tissue from one soil layer to another.

Soil-to-root conductance is the result of two resistances in series, first across the soil-root interface and then through the root tissue. The root tissue conductance is defined as the maximum conductance multiplied by the percent of maximum conductance, as calculated by the sigmoidal vulnerability curve. The maximum conductance is defined as the maximum root-tissue conductivity (PFT parameter) divided by the length of the conducting path, which is taken to be the soil layer depth plus lateral root length.

The soil-root interface conductance is defined as the soil conductivity divided by the conducting length from soil to root. The soil conductivity varies by soil layer and is calculated based on soil potential and soil properties, via the Brooks-Corey theory. The conducting length is determined from the characteristic root spacing (section 2.11.1).

$$q_{3,i} = k_{3,i} \cdot RAI \cdot (\psi_{soil,i} - \psi_{root} + \Delta\psi_{z,i}) \quad (2.583)$$

$$RAI = (LAI + SAI) \cdot r_i \cdot f_{root-leaf} \quad (2.584)$$

$$k_{3,i} = \frac{k_{r,i} \cdot k_{s,i}}{k_{r,i} + k_{s,i}} \quad (2.585)$$

$$k_{r,i} = \frac{k_{3,max}}{z_{3,i}} \cdot 2^{-\left(\frac{\psi_{soil,i}}{p50_3}\right)^{ck}} \quad (2.586)$$

$$k_{s,i} = \frac{k_{soil,i}}{dx_{root,i}} \quad (2.587)$$

Variables:

$q_{3,i}$ = flux of water (mmH₂O/s) from soil layer i to root

$\Delta\psi_{z,i}$ = change in gravitational potential from soil layer i to surface (mmH₂O)

LAI = total leaf area index (m²/m²)

SAI = stem area index (m²/m²)

$\psi_{soil,i}$ = water potential in soil layer i (mmH₂O)

ψ_{root} = root water potential (mmH₂O)

$z_{3,i}$ = length of root tissue conducting path = soil layer depth + root lateral length (m)

r_i = root fraction in soil layer i (-)

$k_{soil,i}$ = Brooks-Corey soil conductivity in soil layer i (m/s)

Parameters:

$f_{root-leaf}$ = root-to-shoot ratio (-)

$p50_3$ = water potential at 50% loss of root tissue conductance (mmH₂O)

ck = shape-fitting parameter for vulnerability curve (-)

Plant Water Demand

Plant water demand depends on stomatal conductance, which is described in section 2.9.3. Here we describe the influence of PHS and the coupling of vegetation water demand and supply. PHS models vegetation water demand as transpiration attenuated by a transpiration loss function based on leaf water potential. Sunlit leaf transpiration is modeled as the maximum sunlit leaf transpiration multiplied by the percent of maximum transpiration as modeled by the sigmoidal loss function. The same follows for shaded leaf transpiration. Maximum stomatal conductance is calculated from the Medlyn model ([Medlyn et al. 2011](#)) absent water stress and used to calculate the maximum transpiration (see section 2.5.3). Water stress is calculated as the ratio of attenuated stomatal conductance to maximum stomatal conductance. Water stress is calculated with distinct values for sunlit and shaded leaves. Vegetation water stress is calculated based on leaf water potential and is used to attenuate photosynthesis (see section 2.9.4)

$$E_{sun} = E_{sun,max} \cdot 2^{-\left(\frac{\psi_{sunleaf}}{p50_e}\right)^{c_k}} \quad (2.588)$$

$$E_{shade} = E_{shade,max} \cdot 2^{-\left(\frac{\psi_{shadeleaf}}{p50_e}\right)^{c_k}} \quad (2.589)$$

$$\beta_{t,sun} = \frac{g_{s,sun}}{g_{s,sun,\beta_t=1}} \quad (2.590)$$

$$\beta_{t,shade} = \frac{g_{s,shade}}{g_{s,shade,\beta_t=1}} \quad (2.591)$$

E_{sun} = sunlit leaf transpiration (mm/s)

E_{shade} = shaded leaf transpiration (mm/s)

$E_{sun,max}$ = sunlit leaf transpiration absent water stress (mm/s)

$E_{shade,max}$ = shaded leaf transpiration absent water stress (mm/s)

$\psi_{sunleaf}$ = sunlit leaf water potential (mmH₂O)

$\psi_{shadeleaf}$ = shaded leaf water potential (mmH₂O)

$\beta_{t,sun}$ = sunlit transpiration water stress (-)

$\beta_{t,shade}$ = shaded transpiration water stress (-)

$g_{s,sun}$ = stomatal conductance of water corresponding to E_{sun}

$g_{s,shade}$ = stomatal conductance of water corresponding to E_{shade}

$g_{s,sun,max}$ = stomatal conductance of water corresponding to $E_{sun,max}$

$g_{s,shade,max}$ = stomatal conductance of water corresponding to $E_{shade,max}$

Vegetation Water Potential

Both plant water supply and demand are functions of vegetation water potential. PHS explicitly models root, stem, shaded leaf, and sunlit leaf water potential at each timestep. PHS iterates to find the vegetation water potential ψ (vector) that satisfies continuity between the non-linear vegetation water supply and demand (equations (2.577), (2.578), (2.581), (2.583), (2.588), (2.589)).

$$\psi = [\psi_{sunleaf}, \psi_{shadeleaf}, \psi_{stem}, \psi_{root}] \quad (2.592)$$

$$\begin{aligned}
E_{sun} &= q_{1a} \\
E_{shade} &= q_{1b} \\
E_{sun} + E_{shade} &= q_{1a} + q_{1b} \\
&= q_2 \\
&= \sum_{i=1}^{nlevsoi} q_{3,i}
\end{aligned} \tag{2.593}$$

PHS finds the water potentials that match supply and demand. In the plant water transport equations (2.593), the demand terms (left-hand side) are decreasing functions of absolute leaf water potential. As absolute leaf water potential becomes larger, water stress increases, causing a decrease in transpiration demand. The supply terms (right-hand side) are increasing functions of absolute leaf water potential. As absolute leaf water potential becomes larger, the gradients in water potential increase, causing an increase in vegetation water supply. PHS takes a Newton's method approach to iteratively solve for the vegetation water potentials that satisfy continuity (2.593).

Numerical Implementation

The four plant water potential nodes are (ψ_{root} , ψ_{xylem} , $\psi_{shadefleaf}$, $\psi_{sunleaf}$). The fluxes between each pair of nodes are labeled in Figure 1. E_{sun} and E_{sha} are the transpiration from sunlit and shaded leaves, respectively. We use the circuit-analog model to calculate the vegetation water potential (ψ) for the four plant nodes, forced by soil matric potential and unstressed transpiration. The unstressed transpiration is acquired by running the photosynthesis model with $\beta_t = 1$. The unstressed transpiration flux is attenuated based on the leaf-level vegetation water potential. Using the attenuated transpiration, we solve for $g_{s,stressed}$ and output $\beta_t = \frac{g_{s,stressed}}{g_{s,unstressed}}$.

The continuity of water flow through the system yields four equations

$$\begin{aligned}
E_{sun} &= q_{1a} \\
E_{shade} &= q_{1b} \\
q_{1a} + q_{1b} &= q_2 \\
q_2 &= \sum_{i=1}^{nlevsoi} q_{3,i}
\end{aligned} \tag{2.594}$$

We seek the set of vegetation water potential values,

$$\psi = \begin{bmatrix} \psi_{sunleaf} \\ \psi_{shadefleaf} \\ \psi_{stem} \\ \psi_{root} \end{bmatrix} \tag{2.595}$$

that satisfies these equations, as forced by the soil moisture and atmospheric state. Each flux on the schematic can be represented in terms of the relevant water potentials. Defining the transpiration fluxes:

$$\begin{aligned}
E_{sun} &= E_{sun,max} \cdot 2^{-\left(\frac{\psi_{sunleaf}}{p50_e}\right)^{c_k}} \\
E_{shade} &= E_{shade,max} \cdot 2^{-\left(\frac{\psi_{shadefleaf}}{p50_e}\right)^{c_k}}
\end{aligned} \tag{2.596}$$

Defining the water supply fluxes:

$$\begin{aligned}
 q_{1a} &= k_{1a,max} \cdot 2^{-\left(\frac{\psi_{stem}}{p50_1}\right)^{c_k}} \cdot LAI_{sun} \cdot (\psi_{stem} - \psi_{sunleaf}) \\
 q_{1b} &= k_{1b,max} \cdot 2^{-\left(\frac{\psi_{stem}}{p50_1}\right)^{c_k}} \cdot LAI_{shade} \cdot (\psi_{stem} - \psi_{shadleaf}) \\
 q_2 &= \frac{k_{2,max}}{z_2} \cdot 2^{-\left(\frac{\psi_{root}}{p50_2}\right)^{c_k}} \cdot SAI \cdot (\psi_{root} - \psi_{stem} - \Delta\psi_z) \\
 q_{soil} &= \sum_{i=1}^{nlevsoi} q_{3,i} = \sum_{i=1}^{nlevsoi} k_{3,i} \cdot RAI \cdot (\psi_{soil,i} - \psi_{root} + \Delta\psi_{z,i})
 \end{aligned} \tag{2.597}$$

We're looking to find the vector ψ that fits with soil and atmospheric forcings while satisfying water flow continuity. Due to the model non-linearity, we use a linearized explicit approach, iterating with Newton's method. The initial guess is the solution for ψ (vector) from the previous time step. The general framework, from iteration m to $m+1$ is:

$$\begin{aligned}
 q^{m+1} &= q^m + \frac{\delta q}{\delta \psi} \Delta\psi \\
 \psi^{m+1} &= \psi^m + \Delta\psi
 \end{aligned} \tag{2.598}$$

So for our first flux balance equation, at iteration $m+1$, we have:

$$E_{sun}^{m+1} = q_{1a}^{m+1} \tag{2.599}$$

Which can be linearized to:

$$E_{sun}^m + \frac{\delta E_{sun}}{\delta \psi} \Delta\psi = q_{1a}^m + \frac{\delta q_{1a}}{\delta \psi} \Delta\psi \tag{2.600}$$

And rearranged to be:

$$\frac{\delta q_{1a}}{\delta \psi} \Delta\psi - \frac{\delta E_{sun}}{\delta \psi} \Delta\psi = E_{sun}^m - q_{1a}^m \tag{2.601}$$

And for the other 3 flux balance equations:

$$\begin{aligned}
 \frac{\delta q_{1b}}{\delta \psi} \Delta\psi - \frac{\delta E_{sha}}{\delta \psi} \Delta\psi &= E_{sha}^m - q_{1b}^m \\
 \frac{\delta q_2}{\delta \psi} \Delta\psi - \frac{\delta q_{1a}}{\delta \psi} \Delta\psi - \frac{\delta q_{1b}}{\delta \psi} \Delta\psi &= q_{1a}^m + q_{1b}^m - q_2^m \\
 \frac{\delta q_{soil}}{\delta \psi} \Delta\psi - \frac{\delta q_2}{\delta \psi} \Delta\psi &= q_2^m - q_{soil}^m
 \end{aligned} \tag{2.602}$$

Putting all four together in matrix form:

$$\begin{bmatrix} \frac{\delta q_{1a}}{\delta \psi} - \frac{\delta E_{sun}}{\delta \psi} \\ \frac{\delta q_{1b}}{\delta \psi} - \frac{\delta E_{sha}}{\delta \psi} \\ \frac{\delta q_2}{\delta \psi} - \frac{\delta q_{1a}}{\delta \psi} - \frac{\delta q_{1b}}{\delta \psi} \\ \frac{\delta q_{soil}}{\delta \psi} - \frac{\delta q_2}{\delta \psi} \end{bmatrix} \Delta\psi = \begin{bmatrix} E_{sun}^m - q_{1a}^m \\ E_{sha}^m - q_{1b}^m \\ q_{1a}^m + q_{1b}^m - q_2^m \\ q_2^m - q_{soil}^m \end{bmatrix} \tag{2.603}$$

Now to expand the left-hand side, from generic ψ to all four plant water potential nodes, noting that many derivatives are zero (e.g. $\frac{\delta E_{sun}}{\delta \psi_{sha}} = 0$)

Introducing the notation: $A\Delta\psi = b$

$$\Delta\psi = \begin{bmatrix} \Delta\psi_{sunleaf} \\ \Delta\psi_{shadleaf} \\ \Delta\psi_{stem} \\ \Delta\psi_{root} \end{bmatrix} \quad (2.604)$$

$$A = \begin{bmatrix} \frac{\delta q_{1a}}{\delta\psi_{sun}} - \frac{\delta E_{sun}}{\delta\psi_{sun}} & 0 & \frac{\delta q_{1a}}{\delta\psi_{stem}} & 0 \\ 0 & \frac{\delta q_{1b}}{\delta\psi_{sha}} - \frac{\delta E_{sha}}{\delta\psi_{sha}} & \frac{\delta q_{1b}}{\delta\psi_{stem}} & 0 \\ -\frac{\delta q_{1a}}{\delta\psi_{sun}} & -\frac{\delta q_{1b}}{\delta\psi_{sha}} & \frac{\delta q_2}{\delta\psi_{stem}} - \frac{\delta q_{1a}}{\delta\psi_{stem}} - \frac{\delta q_{1b}}{\delta\psi_{stem}} & \frac{\delta q_2}{\delta\psi_{root}} \\ 0 & 0 & -\frac{\delta q_2}{\delta\psi_{stem}} & \frac{\delta q_{soil}}{\delta\psi_{root}} - \frac{\delta q_2}{\delta\psi_{root}} \end{bmatrix} \quad (2.605)$$

$$b = \begin{bmatrix} E_{sun}^m - q_{b1}^m \\ E_{sha}^m - q_{b2}^m \\ q_{b1}^m + q_{b2}^m - q_{stem}^m \\ q_{stem}^m - q_{soil}^m \end{bmatrix} \quad (2.606)$$

Now we compute all the entries for A and b based on the soil moisture and maximum transpiration forcings and can solve to find:

$$\Delta\psi = A^{-1}b \quad (2.607)$$

$$\psi_{m+1} = \psi_m + \Delta\psi \quad (2.608)$$

We iterate until $b \rightarrow 0$, signifying water flux balance through the system. The result is a final set of water potentials (ψ_{root} , ψ_{xylem} , $\psi_{shadleaf}$, $\psi_{sunleaf}$) satisfying non-divergent water flux through the system. The magnitude of the water flux is driven by soil matric potential and unstressed ($\beta_t = 1$) transpiration.

We use the transpiration solution (corresponding to the final solution for ψ) to compute stomatal conductance. The stomatal conductance is then used to compute β_t .

$$\beta_{t,sun} = \frac{g_{s,sun}}{g_{s,sun,\beta_t=1}} \quad (2.609)$$

$$\beta_{t,shade} = \frac{g_{s,shade}}{g_{s,shade,\beta_t=1}} \quad (2.610)$$

The β_t values are used in the Photosynthesis module (see section 2.9.4) to apply water stress. The solution for ψ is saved as a new variable (vegetation water potential) and is indicative of plant water status. The soil-to-root fluxes ($q_{3,1}, q_{3,2}, \dots, q_{3,n}$) are used as the soil transpiration sink in the Richards' equation subsurface flow equations (see section 2.7.3).

Flow Diagram of Leaf Flux Calculations:

PHS runs nested in the loop that solves for sensible and latent heat fluxes and temperature for vegetated surfaces (see section 2.5.3). The scheme iterates for convergence of leaf temperature (T_l), transpiration water stress (β_t), and intercellular CO₂ concentration (c_i). PHS is forced by maximum transpiration (absent water stress, $\beta_t = 1$), whereby

we first solve for assimilation, stomatal conductance, and intercellular CO₂ with $\beta_{t,sun}$ and $\beta_{t,shade}$ both set to 1. This involves iterating to convergence of c_i (see section 2.9.4).

Next, using the solutions for $E_{sun,max}$ and $E_{shade,max}$, PHS solves for ψ , $\beta_{t,sun}$, and $\beta_{t,shade}$. The values for $\beta_{t,sun}$, and $\beta_{t,shade}$ are inputs to the photosynthesis routine, which now solves for attenuated photosynthesis and stomatal conductance (reflecting water stress). Again this involves iterating to convergence of c_i . Non-linearities between β_t and transpiration require also iterating to convergence of β_t . The outermost level of iteration works towards convergence of leaf temperature, reflecting leaf surface energy balance.

2.12 Lake Model

The lake model, denoted the *Lake, Ice, Snow, and Sediment Simulator* (LISSS), is from [Subin et al. \(2012a\)](#). It includes extensive modifications to the lake code of [Zeng et al. \(2002\)](#) used in CLM versions 2 through 4, which utilized concepts from the lake models of [Bonan \(1996\)](#), [Henderson-Sellers \(1985\)](#), [Henderson-Sellers \(1986\)](#), [Hostetler and Bartlein \(1990\)](#), and the coupled lake-atmosphere model of [Hostetler et al. \(1993\)](#), [Hostetler et al. \(1993\)](#). Lakes have spatially variable depth prescribed in the surface data (section [External Data](#)); the surface data optionally includes lake optical extinction coefficient and horizontal fetch, currently only used for site simulations. Lake physics includes freezing and thawing in the lake body, resolved snow layers, and “soil” and bedrock layers below the lake body. Temperatures and ice fractions are simulated for $N_{levlak} = 10$ layers (for global simulations) or $N_{levlak} = 25$ (for site simulations) with discretization described in section 2.12.1. Lake albedo is described in section 2.12.3. Lake surface fluxes (section 2.12.4) generally follow the formulations for non-vegetated surfaces, including the calculations of aerodynamic resistances (section 2.5.2); however, the lake surface temperature T_g (representing an infinitesimal interface layer between the top resolved lake layer and the atmosphere) is solved for simultaneously with the surface fluxes. After surface fluxes are evaluated, temperatures are solved simultaneously in the resolved snow layers (if present), the lake body, and the soil and bedrock, using the ground heat flux G as a top boundary condition. Snow, soil, and bedrock models generally follow the formulations for non-vegetated surfaces (Chapter 2.6), with modifications described below.

2.12.1 Vertical Discretization

Currently, there is one lake modeled in each grid cell (with prescribed or assumed depth d , extinction coefficient η , and fetch f), although this could be modified with changes to the CLM subgrid decomposition algorithm in future model versions. As currently implemented, the lake consists of 0-5 snow layers; water and ice layers (10 for global simulations and 25 for site simulations) comprising the “lake body;” 10 “soil” layers; and 5 bedrock layers. Each lake body layer has a fixed water mass (set by the nominal layer thickness and the liquid density), with frozen mass-fraction I a state variable. Resolved snow layers are present if the snow thickness $z_{sno} \geq s_{min}$, where $s_{min} = 4$ cm by default, and is adjusted for model timesteps other than 1800 s in order to maintain numerical stability (section 2.12.6). For global simulations with 10 body layers, the default (50 m lake) body layer thicknesses are given by: Δz_i of 0.1, 1, 2, 3, 4, 5, 7, 7, 10.45, and 10.45 m, with node depths z_i located at the center of each layer (i.e., 0.05, 0.6, 2.1, 4.6, 8.1, 12.6, 18.6, 25.6, 34.325, 44.775 m). For site simulations with 25 layers, the default thicknesses are (m): 0.1 for layer 1; 0.25 for layers 2-5; 0.5 for layers 6-9; 0.75 for layers 10-13; 2 for layers 14-15; 2.5 for layers 16-17; 3.5 for layers 18-21; and 5.225 for layers 22-25. For lakes with depth $d \neq 50$ m and $d \geq 1$ m, the top layer is kept at 10 cm and the other 9 layer thicknesses are adjusted to maintain fixed proportions. For lakes with $d < 1$ m, all layers have equal thickness. Thicknesses of snow, soil, and bedrock layers follow the scheme used over non-vegetated surfaces (Chapter 2.6), with modifications to the snow layer thickness rules to keep snow layers at least as thick as s_{min} (section 2.12.6).

2.12.2 External Data

As discussed in [Subin et al. \(2012a, b\)](#), the Global Lake and Wetland Database ([Lehner and Doll 2004](#)) is currently used to prescribe lake fraction in each land model grid cell, for a total of 2.3 million km⁻². As in [Subin et al. \(2012a,](#)

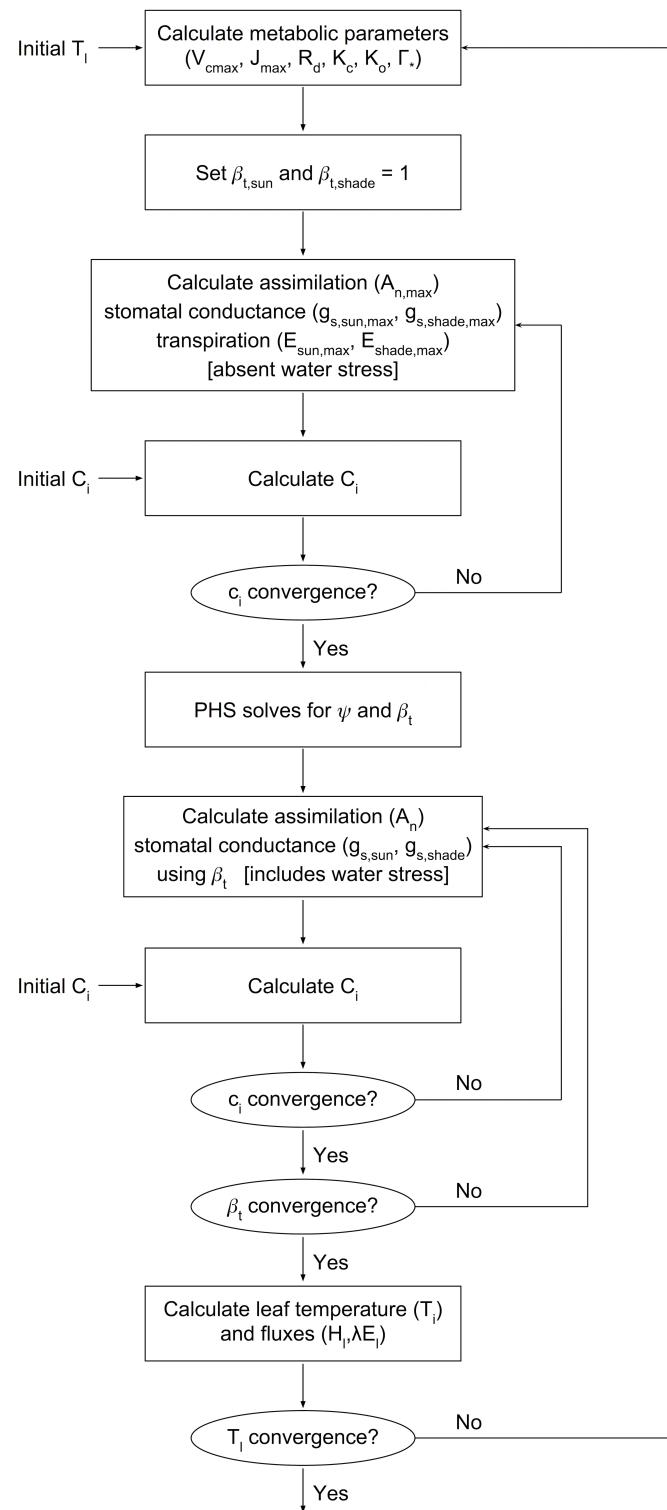


Figure 2.11: Flow diagram of leaf flux calculations

b), the [Kourzeneva et al. \(2012\)](#) global gridded dataset is currently used to estimate a mean lake depth in each grid cell, based on interpolated compilations of geographic information.

2.12.3 Surface Albedo

For direct radiation, the albedo a for lakes with ground temperature T_g (K) above freezing is given by ([Pivovarov, 1972](#))

$$a = \frac{0.5}{\cos z + 0.15} \quad (2.611)$$

where z is the zenith angle. For diffuse radiation, the expression in eq. is integrated over the full sky to yield $a = 0.10$.

For frozen lakes without resolved snow layers, the albedo at cold temperatures a_0 is 0.60 for visible and 0.40 for near infrared radiation. As the temperature at the ice surface, T_g , approaches freezing [T_f (K) ([Table 2.7](#))], the albedo is relaxed towards 0.10 based on [Mironov et al. \(2010\)](#):

$$a = a_0 (1 - x) + 0.10x, x = \exp \left(-95 \frac{T_f - T_g}{T_f} \right) \quad (2.612)$$

where a is restricted to be no less than that given in (2.611).

For frozen lakes with resolved snow layers, the reflectance of the ice surface is fixed at a_0 , and the snow reflectance is calculated as over non-vegetated surfaces ([Chapter 2.3](#)). These two reflectances are combined to obtain the snow-fraction-weighted albedo as in over non-vegetated surfaces ([Chapter 2.3](#)).

2.12.4 Surface Fluxes and Surface Temperature

Surface Properties

The fraction of shortwave radiation absorbed at the surface, β , depends on the lake state. If resolved snow layers are present, then β is set equal to the absorption fraction predicted by the snow-optics submodel ([Chapter 2.3](#)) for the top snow layer. Otherwise, β is set equal to the near infrared fraction of the shortwave radiation reaching the surface simulated by the atmospheric model or atmospheric data model used for offline simulations ([Chapter 2.32](#)). The remainder of the shortwave radiation fraction ($1 - \beta$) is absorbed in the lake body or soil as described in section [2.12.5](#).

The surface roughnesses are functions of the lake state and atmospheric forcing. For frozen lakes ($T_g \leq T_f$) with resolved snow layers, the momentum roughness length $z_{0m} = 2.4 \times 10^{-3}$ m (as over non-vegetated surfaces; [Chapter 2.5](#)), and the scalar roughness lengths (z_{0q} for latent heat; and z_{0h} , for sensible heat) are given by ([Zilitinkevich 1970](#))

$$\begin{aligned} R_0 &= \frac{z_{0m} u_*}{\nu}, \\ z_{0h} &= z_{0q} = z_{0m} \exp \left\{ -0.13 R_0^{0.45} \right\} \end{aligned} \quad (2.613)$$

where R_0 is the near-surface atmospheric roughness Reynolds number, z_{0h} is the roughness length for sensible heat, z_{0q} is the roughness length for latent heat, ν ($\text{m}^2 \text{s}^{-1}$) is the kinematic viscosity of air, and u (m s^{-1}) is the friction velocity in the atmospheric surface layer. For frozen lakes without resolved snow layers, $z_{0m} = 1 \times 10^{-3}$ m ([Subin et al. \(2012a\)](#)), and the scalar roughness lengths are given by .

For unfrozen lakes, z_{0m} is given by ([Subin et al. \(2012a\)](#))

$$z_{0m} = \max \left(\frac{\alpha \nu}{u_*}, C \frac{u_*^2}{g} \right) \quad (2.614)$$

where $\alpha = 0.1$, ν is the kinematic viscosity of air given below, C is the effective Charnock coefficient given below, and g is the acceleration of gravity ([Table 2.7](#)). The kinematic viscosity is given by

$$\nu = \nu_0 \left(\frac{T_g}{T_0} \right)^{1.5} \frac{P_0}{P_{ref}} \quad (2.615)$$

where $\nu_0 = 1.51 \times 10^{-5} \frac{\text{m}^2}{\text{s}}$, $T_0 = 293.15 \text{ K}$, $P_0 = 1.013 \times 10^5 \text{ Pa}$, and P_{ref} is the pressure at the atmospheric reference height. The Charnock coefficient C is a function of the lake fetch F (m), given in the surface data or set to 25 times the lake depth d by default:

$$\begin{aligned} C &= C_{\min} + (C_{\max} - C_{\min}) \exp \{-\min(A, B)\} \\ A &= \left(\frac{Fg}{u^2} \right)^{1/3} / f_c \\ B &= \varepsilon \frac{\sqrt{dg}}{u} \end{aligned} \quad (2.616)$$

where A and B define the fetch- and depth-limitation, respectively; $C_{\min} = 0.01$, $C_{\max} = 0.01$, $\varepsilon = 1$, $f_c = 100$, and u (m s^{-1}) is the atmospheric forcing wind.

Surface Flux Solution

Conservation of energy at the lake surface requires

$$\beta \vec{S}_g - \vec{L}_g - H_g - \lambda E_g - G = 0 \quad (2.617)$$

where \vec{S}_g is the absorbed solar radiation in the lake, β is the fraction absorbed at the surface, \vec{L}_g is the net emitted longwave radiation (+ upwards), H_g is the sensible heat flux (+ upwards), E_g is the water vapor flux (+ upwards), and G is the ground heat flux (+ downwards). All of these fluxes depend implicitly on the temperature at the lake surface T_g . λ converts E_g to an energy flux based on

$$\lambda = \begin{cases} \lambda_{sub} & T_g \leq T_f \\ \lambda_{vap} & T_g > T_f \end{cases}. \quad (2.618)$$

The sensible heat flux (W m^{-2}) is

$$H_g = -\rho_{atm} C_p \frac{(\theta_{atm} - T_g)}{r_{ah}} \quad (2.619)$$

where ρ_{atm} is the density of moist air (kg m^{-3}) (Chapter 2.5), C_p is the specific heat capacity of air ($\text{J kg}^{-1} \text{ K}^{-1}$) (Table 2.7), θ_{atm} is the atmospheric potential temperature (K) (Chapter 2.5), T_g is the lake surface temperature (K) (at an infinitesimal interface just above the top resolved model layer: snow, ice, or water), and r_{ah} is the aerodynamic resistance to sensible heat transfer (s m^{-1}) (section 2.5.1).

The water vapor flux ($\text{kg m}^{-2} \text{ s}^{-1}$) is

$$E_g = -\frac{\rho_{atm} (q_{atm} - q_{sat}^{T_g})}{r_{aw}} \quad (2.620)$$

where q_{atm} is the atmospheric specific humidity (kg kg^{-1}) (section 2.2.3), $q_{sat}^{T_g}$ is the saturated specific humidity (kg kg^{-1}) (section 2.5.5) at the lake surface temperature T_g , and r_{aw} is the aerodynamic resistance to water vapor transfer (s m^{-1}) (section 2.5.1).

The zonal and meridional momentum fluxes are

$$\tau_x = -\rho_{atm} \frac{u_{atm}}{r_{atm}} \quad (2.621)$$

$$\tau_y = -\rho_{atm} \frac{v_{atm}}{r_{atm}} \quad (2.622)$$

where u_{atm} and v_{atm} are the zonal and meridional atmospheric winds (m s^{-1}) (section 2.2.3), and r_{am} is the aerodynamic resistance for momentum (s m^{-1}) (section 2.5.1).

The heat flux into the lake surface G (W m^{-2}) is

$$G = \frac{2\lambda_T}{\Delta z_T} (T_g - T_T) \quad (2.623)$$

where λ_T is the thermal conductivity ($\text{W m}^{-1} \text{K}^{-1}$), Δz_T is the thickness (m), and T_T is the temperature (K) of the top resolved lake layer (snow, ice, or water). The top thermal conductivity λ_T of unfrozen lakes ($T_g > T_f$) includes conductivities due to molecular (λ_{liq}) and eddy (λ_K) diffusivities (section 2.12.5), as evaluated in the top lake layer at the previous timestep, where λ_{liq} is the thermal conductivity of water (Table 2.7). For frozen lakes without resolved snow layers, $\lambda_T = \lambda_{ice}$. When resolved snow layers are present, λ_T is calculated based on the water content, ice content, and thickness of the top snow layer, as for non-vegetated surfaces.

The absorbed solar radiation \vec{S}_g is

$$\vec{S}_g = \sum_{\Lambda} S_{atm} \downarrow_{\Lambda}^{\mu} (1 - \alpha_{g,\Lambda}^{\mu}) + S_{atm} \downarrow_{\Lambda} (1 - \alpha_{g,\Lambda}) \quad (2.624)$$

where $S_{atm} \downarrow_{\Lambda}^{\mu}$ and $S_{atm} \downarrow_{\Lambda}$ are the incident direct beam and diffuse solar fluxes (W m^{-2}) and Λ denotes the visible ($< 0.7 \mu\text{m}$) and near-infrared ($\geq 0.7 \mu\text{m}$) wavebands (section 2.2.3), and $\alpha_{g,\Lambda}^{\mu}$ and $\alpha_{g,\mu}$ are the direct beam and diffuse lake albedos (section 2.12.3).

The net emitted longwave radiation is

$$\vec{L}_g = L_g \uparrow - L_{atm} \downarrow \quad (2.625)$$

where $L_g \uparrow$ is the upward longwave radiation from the surface, $L_{atm} \downarrow$ is the downward atmospheric longwave radiation (section 2.2.3). The upward longwave radiation from the surface is

$$L \uparrow = (1 - \varepsilon_g) L_{atm} \downarrow + \varepsilon_g \sigma (T_g^n)^4 + 4\varepsilon_g \sigma (T_g^n)^3 (T_g^{n+1} - T_g^n) \quad (2.626)$$

where $\varepsilon_g = 0.97$ is the lake surface emissivity, σ is the Stefan-Boltzmann constant ($\text{W m}^{-2} \text{K}^{-4}$) (Table 2.7), and $T_g^{n+1} - T_g^n$ is the difference in lake surface temperature between Newton-Raphson iterations (see below).

The sensible heat H_g , the water vapor flux E_g through its dependence on the saturated specific humidity, the net longwave radiation \vec{L}_g , and the ground heat flux G , all depend on the lake surface temperature T_g . Newton-Raphson iteration is applied to solve for T_g and the surface fluxes as

$$\Delta T_g = \frac{\beta \vec{S}_g - \vec{L}_g - H_g - \lambda E_g - G}{\frac{\partial \vec{L}_g}{\partial T_g} + \frac{\partial H_g}{\partial T_g} + \frac{\partial \lambda E_g}{\partial T_g} + \frac{\partial G}{\partial T_g}} \quad (2.627)$$

where $\Delta T_g = T_g^{n+1} - T_g^n$ and the subscript “n” indicates the iteration. Therefore, the surface temperature T_g^{n+1} can be written as

$$T_g^{n+1} = \frac{\beta \vec{S}_g - \vec{L}_g - H_g - \lambda E_g - G + T_g^n \left(\frac{\partial \vec{L}_g}{\partial T_g} + \frac{\partial H_g}{\partial T_g} + \frac{\partial \lambda E_g}{\partial T_g} + \frac{\partial G}{\partial T_g} \right)}{\frac{\partial \vec{L}_g}{\partial T_g} + \frac{\partial H_g}{\partial T_g} + \frac{\partial \lambda E_g}{\partial T_g} + \frac{\partial G}{\partial T_g}} \quad (2.628)$$

where the partial derivatives are

$$\frac{\partial \vec{L}_g}{\partial T_g} = 4\varepsilon_g \sigma (T_g^n)^3, \quad (2.629)$$

$$\frac{\partial H_g}{\partial T_g} = \frac{\rho_{atm} C_p}{r_{ah}}, \quad (2.630)$$

$$\frac{\partial \lambda E_g}{\partial T_g} = \frac{\lambda \rho_{atm}}{r_{aw}} \frac{dq_{sat}^{T_g}}{dT_g}, \quad (2.631)$$

$$\frac{\partial G}{\partial T_g} = \frac{2\lambda_T}{\Delta z_T}. \quad (2.632)$$

The fluxes of momentum, sensible heat, and water vapor are solved for simultaneously with lake surface temperature as follows. The stability-related equations are the same as for non-vegetated surfaces (section 2.5.2), except that the surface roughnesses are here (weakly varying) functions of the friction velocity u . To begin, z_{0m} is set based on the value calculated for the last timestep (for $T_g > T_f$) or based on the values in section 2.12.4 (otherwise), and the scalar roughness lengths are set based on the relationships in section 2.12.4.

1. An initial guess for the wind speed V_a including the convective velocity U_c is obtained from (2.133) assuming an initial convective velocity $U_c = 0 \text{ m s}^{-1}$ for stable conditions ($\theta_{v, atm} - \theta_{v, s} \geq 0$ as evaluated from (2.159)) and $U_c = 0.5$ for unstable conditions ($\theta_{v, atm} - \theta_{v, s} < 0$).
2. An initial guess for the Monin-Obukhov length L is obtained from the bulk Richardson number using (2.155) and (2.157).
3. The following system of equations is iterated four times:
4. Heat of vaporization / sublimation λ ((2.618))
5. Thermal conductivity λ_T (above)
6. Friction velocity u ((2.141), (2.142), (2.143), (2.144))
7. Potential temperature scale θ ((2.146), (2.147), (2.148), (2.149))
8. Humidity scale q ((2.150), (2.151), (2.152), (2.153))
9. Aerodynamic resistances r_{am} , r_{ah} , and r_{aw} ((2.164), (2.165), (2.166))
10. Lake surface temperature T_g^{n+1} ((2.628))
11. Heat of vaporization / sublimation λ ((2.618))
12. Sensible heat flux H_g is updated for T_g^{n+1} ((2.619))
13. Water vapor flux E_g is updated for T_g^{n+1} as

$$E_g = -\frac{\rho_{atm}}{r_{aw}} \left[q_{atm} - q_{sat}^{T_g} - \frac{\partial q_{sat}^{T_g}}{\partial T_g} (T_g^{n+1} - T_g^n) \right] \quad (2.633)$$

where the last term on the right side of equation is the change in saturated specific humidity due to the change in T_g between iterations.

1. Saturated specific humidity $q_{sat}^{T_g}$ and its derivative $\frac{dq_{sat}^{T_g}}{dT_g}$ are updated for T_g^{n+1} (section 2.5.1).
2. Virtual potential temperature scale θ_v ((2.126))
3. Wind speed including the convective velocity, V_a ((2.133))
4. Monin-Obukhov length L ((2.158))
5. Roughness lengths ((2.613), (2.614)).

Once the four iterations for lake surface temperature have been yielded a tentative solution T_g' , several restrictions are imposed in order to maintain consistency with the top lake model layer temperature T_T (*Subin et al. (2012a)*).

- 1) $T_T \leq T_f < T_g' \Rightarrow T_g = T_f,$
 - 2) $T_T > T_g' > T_m \Rightarrow T_g = T_T,$
 - 3) $T_m > T_g' > T_T > T_f \Rightarrow T_g = T_T$
- (2.634)

where T_m is the temperature of maximum liquid water density, 3.85° C (*Hostetler and Bartlein (1990)*). The first condition requires that, if there is any snow or ice present, the surface temperature is restricted to be less than or equal to freezing. The second and third conditions maintain convective stability in the top lake layer.

If eq. XXX is applied, the turbulent fluxes H_g and E_g are re-evaluated. The emitted longwave radiation and the momentum fluxes are re-evaluated in any case. The final ground heat flux G is calculated from the residual of the energy balance eq. XXX in order to precisely conserve energy. XXX This ground heat flux is taken as a prescribed flux boundary condition for the lake temperature solution (section 2.12.5). An energy balance check is included at each timestep to insure that eq. XXX is obeyed to within 0.1 W m⁻².

2.12.5 Lake Temperature

Introduction

The (optional-) snow, lake body (water and/or ice), soil, and bedrock system is unified for the lake temperature solution. The governing equation, similar to that for the snow-soil-bedrock system for vegetated land units (Chapter 2.6), is

$$\tilde{c}_v \frac{\partial T}{\partial t} = \frac{\partial}{\partial z} \left(\tau \frac{\partial T}{\partial z} \right) - \frac{d\phi}{dz} \quad (2.635)$$

where \tilde{c}_v is the volumetric heat capacity (J m⁻³ K⁻¹), t is time (s), T is the temperature (K), τ is the thermal conductivity (W m⁻¹ K⁻¹), and ϕ is the solar radiation (W m⁻²) penetrating to depth z (m). The system is discretized into N layers, where

$$N = n_{sno} + N_{levlak} + N_{levgrnd}, \quad (2.636)$$

n_{sno} is the number of actively modeled snow layers at the current timestep (Chapter 2.8), and $N_{levgrnd}$ is as for vegetated land units (Chapter 2.6). Energy is conserved as

$$\frac{d}{dt} \sum_{j=1}^N [\tilde{c}_{v,j}(t) (T_j - T_f) + L_j(t)] \Delta z_j = G + (1 - \beta) \vec{S}_g \quad (2.637)$$

where $\tilde{c}_{v,j}(t)$ is the volumetric heat capacity of the j th layer (section 2.12.5), $L_j(t)$ is the latent heat of fusion per unit volume of the j th layer (proportional to the mass of liquid water present), and the right-hand side represents the net influx of energy to the lake system. Note that $\tilde{c}_{v,j}(t)$ can only change due to phase change (except for changing snow layer mass, which, apart from energy required to melt snow, represents an untracked energy flux in the land model, along with advected energy associated with water flows in general), and this is restricted to occur at $T_j = T_f$ in the snow-lake-soil system, allowing eq. to be precisely enforced and justifying the exclusion of $c_{v,j}$ from the time derivative in eq. .

Overview of Changes from CLM4

Thermal conductivities include additional eddy diffusivity, beyond the *Hostetler and Bartlein (1990)* formulation, due to unresolved processes (*Fang and Stefan 1996; Subin et al. (2012a)*). Lake water is now allowed to freeze by an arbitrary fraction for each layer, which releases latent heat and changes thermal properties. Convective mixing occurs for all lakes, even if frozen. Soil and bedrock are included beneath the lake. The full snow model is used if the snow

thickness exceeds a threshold; if there are resolved snow layers, radiation transfer is predicted by the snow-optics submodel (Chapter 2.3), and the remaining radiation penetrating the bottom snow layer is absorbed in the top layer of lake ice; conversely, if there are no snow layers, the solar radiation penetrating the bottom lake layer is absorbed in the top soil layer. The lakes have variable depth, and all physics is assumed valid for arbitrary depth, except for a depth-dependent enhanced mixing (section 2.12.5). Finally, a previous sign error in the calculation of eddy diffusivity (specifically, the Brunt-Väisälä frequency term; eq.) was corrected.

Boundary Conditions

The top boundary condition, imposed at the top modeled layer $i = j_{top}$, where $j_{top} = -n_{sno} + 1$, is the downwards surface flux G defined by the energy flux residual during the surface temperature solution (section 2.12.5). The bottom boundary condition, imposed at $i = N_{levlak} + N_{levgrnd}$, is zero flux. The 2-m windspeed u_2 (m s^{-1}) is used in the calculation of eddy diffusivity:

$$u_2 = \frac{u_*}{k} \ln \left(\frac{2}{z_{0m}} \right) \geq 0.1. \quad (2.638)$$

where u_* is the friction velocity calculated in section 2.12.5 and k is the von Karman constant (Table 2.7).

Eddy Diffusivity and Thermal Conductivities

The total eddy diffusivity K_W ($\text{m}^2 \text{s}^{-1}$) for liquid water in the lake body is given by (Subin et al. (2012a))

$$K_W = m_d (\kappa_e + K_{ed} + \kappa_m) \quad (2.639)$$

where κ_e is due to wind-driven eddies (Hostetler and Bartlein (1990)), K_{ed} is a modest enhanced diffusivity intended to represent unresolved mixing processes (Fang and Stefan 1996), $\kappa_m = \frac{\lambda_{liq}}{c_{liq}\rho_{liq}}$ is the molecular diffusivity of water (given by the ratio of its thermal conductivity ($\text{W m}^{-1} \text{K}^{-1}$) to the product of its heat capacity ($\text{J kg}^{-1} \text{K}^{-1}$) and density (kg m^{-3}), values given in Table 2.7), and m_d (unitless) is a factor which increases the overall diffusivity for large lakes, intended to represent 3-dimensional mixing processes such as caused by horizontal temperature gradients. As currently implemented,

$$m_d = \begin{cases} 1, & d < 25\text{m} \\ 10, & d \geq 25\text{m} \end{cases} \quad (2.640)$$

where d is the lake depth.

The wind-driven eddy diffusion coefficient $\kappa_{e,i}$ ($\text{m}^2 \text{s}^{-1}$) for layers $1 \leq i \leq N_{levlak}$ is

$$\kappa_{e,i} = \begin{cases} \frac{k w^* z_i}{P_0(1+37Ri^2)} \exp(-k^* z_i) & T_g > T_f \\ 0 & T_g \leq T_f \end{cases} \quad (2.641)$$

where $P_0 = 1$ is the neutral value of the turbulent Prandtl number, z_i is the node depth (m), the surface friction velocity (m s^{-1}) is $w^* = 0.0012u_2$, and k^* varies with latitude ϕ as $k^* = 6.6u_2^{-1.84}\sqrt{|\sin \phi|}$. For the bottom layer, $\kappa_{e,N_{levlak}} = \kappa_{e,N_{levlak}-1}$. As in Hostetler and Bartlein (1990), the 2-m wind speed u_2 (m s^{-1}) (eq.) is used to evaluate w^* and k^* rather than the 10-m wind used by Henderson-Sellers (1985).

The Richardson number is

$$R_i = \frac{-1 + \sqrt{1 + \frac{40N^2 k^2 z_i^2}{w^{*2} \exp(-2k^* z_i)}}}{20} \quad (2.642)$$

where

$$N^2 = \frac{g}{\rho_i} \frac{\partial \rho}{\partial z} \quad (2.643)$$

and g is the acceleration due to gravity (m s^{-2}) (Table 2.7), ρ_i is the density of water (kg m^{-3}), and $\frac{\partial \rho}{\partial z}$ is approximated as $\frac{\rho_{i+1}-\rho_i}{z_{i+1}-z_i}$. Note that because here, z is increasing downwards (unlike in [Hostetler and Bartlein \(1990\)](#)), eq. contains no negative sign; this is a correction from CLM4. The density of water is ([Hostetler and Bartlein \(1990\)](#))

$$\rho_i = 1000 \left(1 - 1.9549 \times 10^{-5} |T_i - 277|^{1.68} \right). \quad (2.644)$$

The enhanced diffusivity K_{ed} is given by ([Fang and Stefan 1996](#))

$$K_{ed} = 1.04 \times 10^{-8} (N^2)^{-0.43}, N^2 \geq 7.5 \times 10^{-5} \text{s}^2 \quad (2.645)$$

where N^2 is calculated as in eq. except for the minimum value imposed in .

The thermal conductivity for the liquid water portion of lake body layer i , $\tau_{liq,i}$ ($\text{W m}^{-1} \text{K}^{-1}$) is given by

$$\tau_{liq,i} = K_W c_{liq} \rho_{liq}. \quad (2.646)$$

The thermal conductivity of the ice portion of lake body layer i , $\tau_{ice,eff}$ ($\text{W m}^{-1} \text{K}^{-1}$), is constant among layers, and is given by

$$\tau_{ice,eff} = \tau_{ice} \frac{\rho_{ice}}{\rho_{liq}} \quad (2.647)$$

where τ_{ice} (Table 2.7) is the nominal thermal conductivity of ice: $\tau_{ice,eff}$ is adjusted for the fact that the nominal model layer thicknesses remain constant even while the physical ice thickness exceeds the water thickness.

The overall thermal conductivity τ_i for layer i with ice mass-fraction I_i is the harmonic mean of the liquid and water fractions, assuming that they will be physically vertically stacked, and is given by

$$\tau_i = \frac{\tau_{ice,eff} \tau_{liq,i}}{\tau_{liq,i} I_i + \tau_{ice} (1 - I_i)}. \quad (2.648)$$

The thermal conductivity of snow, soil, and bedrock layers above and below the lake, respectively, are computed identically to those for vegetated land units (Chapter 2.6), except for the adjustment of thermal conductivity for frost heave or excess ice ([Subin et al., 2012a, Supporting Information](#)).

Radiation Penetration

If there are no resolved snow layers, the surface absorption fraction β is set according to the near-infrared fraction simulated by the atmospheric model. This is apportioned to the surface energy budget (section 2.12.4), and thus no additional radiation is absorbed in the top z_a (currently 0.6 m) of unfrozen lakes, for which the light extinction coefficient η (m^{-1}) varies between lake columns (eq.). For frozen lakes ($T_g \leq T_f$), the remaining $(1 - \beta) \vec{S}_g$ fraction of surface absorbed radiation that is not apportioned to the surface energy budget is absorbed in the top lake body layer. This is a simplification, as lake ice is partially transparent. If there are resolved snow layers, then the snow optics submodel (Chapter 2.3) is used to calculate the snow layer absorption (except for the absorption predicted for the top layer by the snow optics submodel, which is assigned to the surface energy budget), with the remainder penetrating snow layers absorbed in the top lake body ice layer.

For unfrozen lakes, the solar radiation remaining at depth $z > z_a$ in the lake body is given by

$$\phi = (1 - \beta \vec{S}_g) \exp \{ -\eta (z - z_a) \}. \quad (2.649)$$

For all lake body layers, the flux absorbed by the layer i , ϕ_i , is

$$\phi_i = (1 - \beta \vec{S}_g) \left[\exp \left\{ -\eta \left(z_i - \frac{\Delta z_i}{2} - z_a \right) \right\} - \exp \left\{ -\eta \left(z_i + \frac{\Delta z_i}{2} - z_a \right) \right\} \right]. \quad (2.650)$$

The argument of each exponent is constrained to be non-negative (so $\phi_i = 0$ for layers contained within z_a). The remaining flux exiting the bottom of layer $i = N_{levlak}$ is absorbed in the top soil layer.

The light extinction coefficient η (m^{-1}), if not provided as external data, is a function of depth d (m) ([Subin et al. \(2012a\)](#)):

$$\eta = 1.1925 d^{-0.424}. \quad (2.651)$$

Heat Capacities

The vertically-integrated heat capacity for each lake layer, $c_{v,i}$ (J m^{-2}) is determined by the mass-weighted average over the heat capacities for the water and ice fractions:

$$c_{v,i} = \Delta z_i \rho_{liq} [c_{liq} (1 - I_i) + c_{ice} I_i]. \quad (2.652)$$

Note that the density of water is used for both ice and water fractions, as the thickness of the layer is fixed.

The total heat capacity $c_{v,i}$ for each soil, snow, and bedrock layer (J m^{-2}) is determined as for vegetated land units (Chapter 2.6), as the sum of the heat capacities for the water, ice, and mineral constituents.

Crank-Nicholson Solution

The solution method for thermal diffusion is similar to that used for soil (Chapter 2.6), except that the lake body layers are sandwiched between the snow and soil layers (section 2.12.5), and radiation flux is absorbed throughout the lake layers. Before solution, layer temperatures T_i (K), thermal conductivities τ_i ($\text{W m}^{-1} \text{K}^{-1}$), heat capacities $c_{v,i}$ (J m^{-2}), and layer and interface depths from all components are transformed into a uniform set of vectors with length $N = n_{sno} + N_{levlak} + N_{levgrnd}$ and consistent units to simplify the solution. Thermal conductivities at layer interfaces are calculated as the harmonic mean of the conductivities of the neighboring layers:

$$\lambda_i = \frac{\tau_i \tau_{i+1} (z_{i+1} - z_i)}{\tau_i (z_{i+1} - \hat{z}_i) + \tau_{i+1} (\hat{z}_i - z_i)}, \quad (2.653)$$

where λ_i is the conductivity at the interface between layer i and layer $i + 1$, z_i is the depth of the node of layer i , and \hat{z}_i is the depth of the interface below layer i . Care is taken at the boundaries between snow and lake and between lake and soil. The governing equation is discretized for each layer as

$$\frac{c_{v,i}}{\Delta t} (T_i^{n+1} - T_i^n) = F_{i-1} - F_i + \phi_i \quad (2.654)$$

where superscripts $n + 1$ and n denote values at the end and beginning of the timestep Δt , respectively, F_i (W m^{-2}) is the downward heat flux at the bottom of layer i , and ϕ_i is the solar radiation absorbed in layer i .

Eq. is solved using the semi-implicit Crank-Nicholson Method, resulting in a tridiagonal system of equations:

$$\begin{aligned} r_i &= a_i T_{i-1}^{n+1} + b_i T_i^{n+1} + c T_{i+1}^{n+1}, \\ a_i &= -0.5 \frac{\Delta t}{c_{v,i}} \frac{\partial F_{i-1}}{\partial T_{i-1}^n}, \\ b_i &= 1 + 0.5 \frac{\Delta t}{c_{v,i}} \left(\frac{\partial F_{i-1}}{\partial T_{i-1}^n} + \frac{\partial F_i}{\partial T_i^n} \right), \\ c_i &= -0.5 \frac{\Delta t}{c_{v,i}} \frac{\partial F_i}{\partial T_i^n}, \\ r_i &= T_i^n + 0.5 \frac{\Delta t}{c_{v,i}} (F_{i-1} - F_i) + \frac{\Delta t}{c_{v,i}} \phi_i. \end{aligned} \quad (2.655)$$

The fluxes F_i are defined as follows: for the top layer, $F_{j_{top}-1} = 2G$; $a_{j_{top}} = 0$, where G is defined as in section 2.12.5 (the factor of 2 merely cancels out the Crank-Nicholson 0.5 in the equation for $r_{j_{top}}$). For the bottom layer, $F_{N_{levlak}+N_{levgrnd}} = 0$. For all other layers:

$$F_i = \lambda_i \frac{T_i^n - T_{i+1}^n}{z_{i+1} - z_i}. \quad (2.656)$$

Phase Change

Phase change in the lake, snow, and soil is done similarly to that done for the soil and snow for vegetated land units (Chapter 2.6), except without the allowance for freezing point depression in soil underlying lakes. After the heat diffusion is calculated, phase change occurs in a given layer if the temperature is below freezing and liquid water remains, or if the temperature is above freezing and ice remains.

If melting occurs, the available energy for melting, Q_{avail} (J m^{-2}), is computed as

$$Q_{avail} = (T_i - T_f) c_{v,i} \quad (2.657)$$

where T_i is the temperature of the layer after thermal diffusion (section 2.12.5), and $c_{v,i}$ is as calculated in section 2.12.5. The mass of melt in the layer M (kg m^{-2}) is given by

$$M = \min \left\{ M_{ice}, \frac{Q_{avail}}{H_{fus}} \right\} \quad (2.658)$$

where H_{fus} (J kg^{-1}) is the latent heat of fusion of water (Table 2.7), and M_{ice} is the mass of ice in the layer: $I_i \rho_{liq} \Delta z_i$ for a lake body layer, or simply the soil / snow ice content state variable (w_{ice}) for a soil / snow layer. The heat remainder, Q_{rem} is given by

$$Q_{rem} = Q_{avail} - M H_{fus}. \quad (2.659)$$

Finally, the mass of ice in the layer M_{ice} is adjusted downwards by M , and the temperature T_i of the layer is adjusted to

$$T_i = T_f + \frac{Q_{rem}}{c'_{v,i}} \quad (2.660)$$

where $c'_{v,i} = c_{v,i} + M (c_{liq} - c_{ice})$.

If freezing occurs, Q_{avail} is again given by but will be negative. The melt M , also negative, is given by

$$M = \max \left\{ -M_{liq}, \frac{Q_{avail}}{H_{fus}} \right\} \quad (2.661)$$

where M_{liq} is the mass of water in the layer: $(1 - I_i) \rho_{liq} \Delta z_i$ for a lake body layer, or the soil / snow water content state variable (w_{liq}). The heat remainder Q_{rem} is given by eq. and will be negative or zero. Finally, M_{liq} is adjusted downwards by $-M$ and the temperature is reset according to eq. .

In the presence of nonzero snow water W_{sno} without resolved snow layers over

an unfrozen top lake layer, the available energy in the top lake layer ($T_1 - T_f$) $c_{v,1}$ is used to melt the snow. Similar to above, W_{sno} is either completely melted and the remainder of heat returned to the top lake layer, or the available heat is exhausted and the top lake layer is set to freezing. The snow thickness is adjusted downwards in proportion to the amount of melt, maintaining constant density.

Convection

Convective mixing is based on [Hostetler et al.'s \(1993, 1994\)](#) coupled lake-atmosphere model, adjusting the lake temperature after diffusion and phase change to maintain a stable density profile. Unfrozen lakes overturn when $\rho_i > \rho_{i+1}$, in which case the layer thickness weighted average temperature for layers 1 to $i + 1$ is applied to layers 1 to $i + 1$ and the densities are updated. This scheme is applied iteratively to layers $1 \leq i < N_{levlak} - 1$. Unstable profiles occurring at the bottom of the lake (i.e., between layers $i = N_{levlak} - 1$ and $i = N_{levlak}$) are treated separately ([Subin et al. \(2012a\)](#)), as occasionally these can be induced by heat expelled from the sediments (not present in the original [Hostetler et al. \(1994\)](#) model). Mixing proceeds from the bottom upward in this case (i.e., first mixing layers $i = N_{levlak} - 1$ and $i = N_{levlak}$, then checking $i = N_{levlak} - 2$ and $i = N_{levlak} - 1$ and mixing down to $i = N_{levlak}$ if needed, and on to the top), so as not to mix in with warmer over-lying layers.

For frozen lakes, this algorithm is generalized to conserve total enthalpy and ice content, and to maintain ice contiguous at the top of the lake. Thus, an additional mixing criterion is added: the presence of ice in a layer that is below a layer which is not completely frozen. When this occurs, these two lake layers and all those above mix. Total enthalpy Q is conserved as

$$Q = \sum_{j=1}^{i+1} \Delta z_j \rho_{liq} (T_j - T_f) [(1 - I_j) c_{liq} + I_j c_{ice}]. \quad (2.662)$$

Once the average ice fraction I_{av} is calculated from

$$I_{av} = \frac{\sum_{j=1}^{i+1} I_j \Delta z_j}{Z_{i+1}}, \quad Z_{i+1} = \sum_{j=1}^{i+1} \Delta z_j, \quad (2.663)$$

the temperatures are calculated. A separate temperature is calculated for the frozen (T_{froz}) and unfrozen (T_{unfr}) fractions of the mixed layers. If the total heat content Q is positive (e.g. some layers will be above freezing), then the extra heat is all assigned to the unfrozen layers, while the fully frozen layers are kept at freezing. Conversely, if $Q < 0$, the heat deficit will all be assigned to the ice, and the liquid layers will be kept at freezing. For the layer that contains both ice and liquid (if present), a weighted average temperature will have to be calculated.

If $Q > 0$, then $T_{froz} = T_f$, and T_{unfr} is given by

$$T_{unfr} = \frac{Q}{\rho_{liq} Z_{i+1} [(1 - I_{av}) c_{liq}]} + T_f. \quad (2.664)$$

If $Q < 0$, then $T_{unfr} = T_f$, and T_{froz} is given by

$$T_{froz} = \frac{Q}{\rho_{liq} Z_{i+1} [I_{av} c_{ice}]} + T_f. \quad (2.665)$$

The ice is lumped together at the top. For each lake layer j from 1 to $i + 1$, the ice fraction and temperature are set as follows, where $Z_j = \sum_{m=1}^j \Delta z_m$:

1. If $Z_j \leq Z_{i+1} I_{av}$, then $I_j = 1$ and $T_j = T_{froz}$.
2. Otherwise, if $Z_{j-1} < Z_{i+1} I_{av}$, then the layer will contain both ice and water. The ice fraction is given by $I_j = \frac{Z_{i+1} I_{av} - Z_{j-1}}{\Delta z_j}$. The temperature is set to conserve the desired heat content that would be present if the layer could have two temperatures, and then dividing by the heat capacity of the layer to yield

$$T_j = \frac{T_{froz} I_j c_{ice} + T_{unfr} (1 - I_j) c_{liq}}{I_j c_{ice} + (1 - I_j) c_{liq}}. \quad (2.666)$$

3. Otherwise, $I_j = 0$ and $T_j = T_{unfr}$.

Energy Conservation

To check energy conservation, the left-hand side of eq. XXX is re-written to yield the total enthalpy of the lake system (J m^{-2}) H_{tot} :

$$H_{tot} = \sum_{i=j_{top}}^{N_{levlak} + N_{levgrnd}} [c_{v,i} (T_i - T_f) + M_{liq,i} H_{fus}] - W_{sno,bulk} H_{fus} \quad (2.667)$$

where $M_{liq,i}$ is the water mass of the i th layer (similar to section 2.12.5), and $W_{sno,bulk}$ is the mass of snow-ice not present in resolved snow layers. This expression is evaluated once at the beginning and once at the end of the timestep (re-evaluating each $c_{v,i}$), and the change is compared with the net surface energy flux to yield the error flux E_{soi} (W m^{-2}):

$$E_{soi} = \frac{\Delta H_{tot}}{\Delta t} - G - \sum_{i=j_{top}}^{N_{levlak} + N_{levgrnd}} \phi_i \quad (2.668)$$

If $|E_{soi}| < 0.1 \text{W m}^{-2}$, it is subtracted from the sensible heat flux and added to G . Otherwise, the model is aborted.

2.12.6 Lake Hydrology

Overview

Hydrology is done similarly to other impervious non-vegetated columns (e.g., glaciers) where snow layers may be resolved but infiltration into the permanent ground is not allowed. The water mass of lake columns is currently maintained constant, aside from overlying snow. The water budget is balanced with q_{rgwl} (eq. ; $\text{kg m}^{-2} \text{s}^{-1}$), a generalized runoff term for impervious land units that may be negative.

There are some modifications to the soil and snow parameterizations as compared with the soil in vegetated land units, or the snow overlying other impervious columns. The soil can freeze or thaw, with the allowance for frost heave (or the initialization of excess ice) (sections 2.12.5 and 2.12.5), but no air-filled pore space is allowed in the soil. To preserve numerical stability in the lake model (which uses a slightly different surface flux algorithm than over other non-vegetated land units), two changes are made to the snow model. First, dew or frost is not allowed to be absorbed by a top snow layer which has become completely melted during the timestep. Second, because occasional instabilities occurred during model testing when the Courant–Friedrichs–Lowy (CFL) condition was violated, due to the explicit time-stepping integration of the surface flux solution, resolved snow layers must be a minimum of $s_{\min} = 4 \text{ cm}$ thick rather than 1 cm when the default timestep of 1800 s is used.

Water Balance

The total water balance of the system is given by

$$\Delta W_{sno} + \sum_{i=1}^{n_{levsoi}} (\Delta w_{liq,i} + \Delta w_{ice,i}) = (q_{rain} + q_{sno} - E_g - q_{rgwl} - q_{snwcp,ice}) \Delta t \quad (2.669)$$

where W_{sno} (kg m^{-2}) is the total mass of snow (both liquid and ice, in resolved snow layers or bulk snow), $w_{liq,i}$ and $w_{ice,i}$ are the masses of water phases (kg m^{-2}) in soil layer i , q_{rain} and q_{sno} are the precipitation forcing from the atmosphere ($\text{kg m}^{-2} \text{s}^{-1}$), $q_{snwcp,ice}$ is the ice runoff associated with snow-capping (below), E_g is the ground evaporation (section 2.12.4), and n_{levsoi} is the number of hydrologically active soil layers (as opposed to dry bedrock layers).

Precipitation, Evaporation, and Runoff

All precipitation reaches the ground, as there is no vegetated fraction. As for other land types, incident snowfall accumulates (with ice mass W_{sno} and thickness z_{sno}) until its thickness exceeds a minimum thickness s_{\min} , at which point a resolved snow layer is initiated, with water, ice, dissolved aerosol, snow-grain radius, etc., state variables tracked by the Snow Hydrology submodel (Chapter 2.8). The density of fresh snow is assigned as for other land types (Chapter 2.8). Solid precipitation is added immediately to the snow, while liquid precipitation is added to snow layers, if they exist, after accounting for dew, frost, and sublimation (below). If z_{sno} exceeds s_{\min} after solid precipitation is added but no snow layers are present, a new snow layer is initiated immediately, and then dew, frost, and sublimation are accounted for. Snow-capping is invoked if the snow depth $z_{sno} > 1000\text{m}$, in which case additional precipitation and frost deposition is added to $q_{snwcp,ice}$.

If there are resolved snow layers, the generalized “evaporation” E_g (i.e., evaporation, dew, frost, and sublimation) is treated as over other land units, except that the allowed evaporation from the ground is unlimited (though the top snow layer cannot lose more water mass than it contains). If there are no resolved snow layers but $W_{sno} > 0$ and $E_g > 0$, sublimation $q_{sub,sno}$ ($\text{kg m}^{-2} \text{s}^{-1}$) will be given by

$$q_{sub,sno} = \min \left\{ E_g, \frac{W_{sno}}{\Delta t} \right\}. \quad (2.670)$$

If $E_g < 0$, $T_g \leq T_f$, and there are no resolved snow layers or the top snow layer is not unfrozen, then the rate of frost production $q_{frost} = |E_g|$. If $E_g < 0$ but the top snow layer has completely thawed during the Phase Change step of

the Lake Temperature solution (section 2.12.5), then frost (or dew) is not allowed to accumulate ($q_{frost} = 0$), to insure that the layer is eliminated by the Snow Hydrology (Chapter 2.8) code. (If $T_g > T_f$, then no snow is present (section 2.12.4), and evaporation or dew deposition is balanced by q_{rgwl} .) The snowpack is updated for frost and sublimation:

$$W_{sno} = W_{sno} + \Delta t (q_{frost} - q_{sub,sno}). \quad (2.671)$$

If there are resolved snow layers, then this update occurs using the Snow Hydrology submodel (Chapter 2.8). Otherwise, the snow ice mass is updated directly, and z_{sno} is adjusted by the same proportion as the snow ice (i.e., maintaining the same density), unless there was no snow before adding the frost, in which case the density is assumed to be 250 kg m^{-3} .

Soil Hydrology

The combined water and ice soil volume fraction in a soil layer θ_i is given by

$$\theta_i = \frac{1}{\Delta z_i} \left(\frac{w_{ice,i}}{\rho_{ice}} + \frac{w_{liq,i}}{\rho_{liq}} \right). \quad (2.672)$$

If $\theta_i < \theta_{sat,i}$, the pore volume fraction at saturation (as may occur when ice melts), then the liquid water mass is adjusted to

$$w_{liq,i} = \left(\theta_{sat,i} \Delta z_i - \frac{w_{ice,i}}{\rho_{ice}} \right) \rho_{liq}. \quad (2.673)$$

Otherwise, if excess ice is melting and $w_{liq,i} > \theta_{sat,i} \rho_{liq} \Delta z_i$, then the water in the layer is reset to

$$w_{liq,i} = \theta_{sat,i} \rho_{liq} \Delta z_i \quad (2.674)$$

This allows excess ice to be initialized (and begin to be lost only after the pore ice is melted, which is realistic if the excess ice is found in heterogeneous chunks) but irreversibly lost when melt occurs.

Modifications to Snow Layer Logic

A thickness difference $z_{lsa} = s_{\min} - \tilde{s}_{\min}$ adjusts the minimum resolved snow layer thickness for lake columns as compared to non-lake columns. The value of z_{lsa} is chosen to satisfy the CFL condition for the model timestep. By default, $\tilde{s}_{\min} = 1 \text{ cm}$ and $s_{\min} = 4 \text{ cm}$. See [Subin et al. \(2012a; including Supporting Information\)](#) for further discussion.

The rules for combining and sub-dividing snow layers (section 2.8.7) are adjusted for lakes to maintain minimum thicknesses of s_{\min} and to increase all target layer thicknesses by z_{lsa} . The rules for combining layers are modified by simply increasing layer thickness thresholds by z_{lsa} . The rules for dividing snow layers are contained in a separate subroutine that is modified for lakes, and is a function of the number of layers and the layer thicknesses. There are two types of operations: (a) subdividing layers in half, and (b) shifting some volume from higher layers to lower layers (without increasing the layer number). For subdivisions of type (a), the thickness thresholds triggering subdivision are increased by $2z_{lsa}$ for lakes. For shifts of type (b), the thickness thresholds triggering the shifts are increased by z_{lsa} . At the end of the modified subroutine, a snow ice and liquid balance check are performed.

In rare instances, resolved snow layers may be present over an unfrozen top lake body layer. In this case, the snow layers may be eliminated if enough heat is present in the top layer to melt the snow: see [Subin et al. \(2012a, Supporting Information\)](#).

2.13 Glaciers

This chapter describes features of CLM that are specific to coupling to an ice sheet model (in the CESM context, this is the Glimmer-CISM model; [Lipscomb and Sacks \(2012\)](#) provide documentation and user's guide for Glimmer-CISM). General information about glacier land units can be found elsewhere in this document (see Chapter 2.2 for an overview).

2.13.1 Overview

CLM is responsible for computing three quantities that are passed to the ice sheet model:

1. Surface mass balance (SMB) – the net annual accumulation/ablation of mass at the upper surface (section 2.13.3)
2. Ground surface temperature, which serves as an upper boundary condition for Glimmer-CISM's temperature calculation
3. Surface topography, which currently is fixed in time, and is provided on CLM's surface dataset

The ice sheet model is typically run at much higher resolution than CLM (e.g., \sim 5 km rather than \sim 100 km). To improve the downscaling from CLM's grid to the ice sheet grid, the glaciated portion of each grid cell is divided into multiple elevation classes (section 10.2). The above quantities are computed separately in each elevation class. Glimmer-CISM then computes high-resolution quantities via horizontal and vertical interpolation.

There are several reasons for computing the SMB in CLM rather than in Glimmer-CISM:

1. It is much cheaper to compute the SMB in CLM for \sim 10 elevation classes than in Glimmer-CISM. For example, suppose we are running CLM at a resolution of \sim 50 km and Glimmer at \sim 5 km. Greenland has dimensions of about 1000 x 2000 km. For CLM we would have $20 \times 40 \times 10 = 8,000$ columns, whereas for Glimmer we would have $200 \times 400 = 80,000$ columns.
2. We can use the sophisticated snow physics parameterization already in CLM instead of implementing a separate scheme for Glimmer-CISM. Any improvements to the CLM are applied to ice sheets automatically.
3. The atmosphere model can respond during runtime to ice-sheet surface changes. As shown by [Pritchard et al. \(2008\)](#), runtime albedo feedback from the ice sheet is critical for simulating ice-sheet retreat on paleoclimate time scales. Without this feedback the atmosphere warms much less, and the retreat is delayed.
4. Mass is more nearly conserved, given that the rate of surface ice growth or melting computed in CLM is equal to the rate seen by the dynamic ice sheet model. (Mass conservation is not exact, however, because of approximations made in interpolating from the CLM grid to the ice-sheet grid.)
5. The improved SMB is available in CLM for all glaciated grid cells (e.g., in the Alps, Rockies, Andes, and Himalayas), not just those which are part of ice sheets.

The current coupling between CLM and Glimmer-CISM is one-way only. That is, CLM sends the SMB and surface temperature to Glimmer-CISM but does not do anything with the fields that are returned. The CLM glacier fraction and surface topography are therefore fixed in time. One-way coupling is reasonable for runs of \sim 100 years or less, in which ice-sheet elevation changes are modest. For longer runs with larger elevation changes, two-way coupling is highly desirable. A two-way coupling scheme is under development.

2.13.2 Multiple elevation class scheme

In the typical operation of CLM, the glacier land unit contains a single column (section 2.2.1). However, when running CESM with an active ice sheet model, the glacier land unit is replaced by a glacier_mec land unit, where "mec" denotes "multiple elevation classes". In most ways, glacier_mec land units behave the same as standard glacier land units. However, each glacier_mec land unit is divided into a user-defined set of columns based on surface elevation. The default is 10 elevation classes whose lower limits are 0, 200, 400, 700, 1000, 1300, 1600, 2000, 2500, and 3000

m. Each column is characterized by a fractional area and surface elevation that are read in during model initialization. Each glacier_mec column within a grid cell has distinct ice and snow temperatures, snow water content, surface fluxes, and SMB.

Glacier_mec columns, like glacier columns, are initialized with a temperature of 250 K. While glacier columns are initialized with a snow liquid water equivalent (LWE) equal to the maximum allowed value of 1 m, glacier_mec columns begin with a snow LWE of 0.5 m so that they will reach their equilibrium mean snow depth sooner. Glacier_mec columns typically require several decades of spin-up to equilibrate with a given climate.

The atmospheric surface temperature, potential temperature, specific humidity, density, and pressure are downscaled from the mean grid cell elevation to the glacier_mec column elevation using a specified lapse rate (typically 6.0 deg/km) and an assumption of uniform relative humidity. At a given time, lower-elevation columns can undergo surface melting while columns at higher elevations remain frozen. This gives a more accurate simulation of summer melting, which is a highly nonlinear function of air temperature. The precipitation rate and radiative fluxes are not currently downscaled, but could be in the future if care were taken to preserve the cell-integrated values.

In contrast to most CLM subgrid units, glacier_mec columns can be active (i.e., have model calculations run there) even if their area is zero. This is done because the ice sheet model may require a SMB even for some grid cells where CLM does not have glacier land units. To allow for this, grid overlap files have been pre-computed. For given resolutions of CLM and Glimmer-CISM, these files identify all land-covered grid cells that overlap any part of the ice sheet grid. In these overlapping cells, glacier_mec columns are defined in all elevation classes. Some columns may have zero area and are called “virtual” columns. These columns do not affect energy exchange between the land and the atmosphere, but are included for potential forcing of Glimmer-CISM.

2.13.3 Computation of the surface mass balance

The SMB of a glacier or ice sheet is the net annual accumulation/ablation of mass at the upper surface. Ablation is defined as the mass of water that runs off to the ocean. Not all the surface meltwater runs off; some of the melt percolates into the snow and refreezes. Accumulation is primarily by snowfall and deposition, and ablation is primarily by melting and evaporation/sublimation. CLM uses a surface-energy-balance (SEB) scheme to compute the SMB. In this scheme, the melting depends on the sum of the radiative, turbulent, and conductive fluxes reaching the surface, as described elsewhere in this document.

CLM has a somewhat unrealistic treatment of accumulation and melting for standard glacier land units. The snow depth is limited to a prescribed depth of 1 m liquid water equivalent, with any additional snow assumed to run off to the ocean. (This amounts to a crude parameterization of iceberg calving.) Snow melting is treated in a realistic fashion, with meltwater percolating downward through snow layers as long as the snow is unsaturated. Once the underlying snow is saturated, any additional meltwater runs off. When glacier ice melts, however, the meltwater is assumed to remain in place until it refreezes. In warm parts of the ice sheet, the meltwater does not refreeze, but stays in place indefinitely.

In the modified glacier_mec columns, the treatment of melting and freezing depends on the logical variable *glc_dyntopo*. This variable controls whether CLM surface topography changes dynamically as the ice sheet evolves (i.e., whether the coupling is one-way or two-way). If *glc_dyntopo* is true, then CLM receives updated topographic information from the ice sheet model. In this case, snow in excess of the prescribed maximum depth is assumed to turn into ice, contributing a positive SMB to the ice sheet model. Melting ice is assumed to run off to the ocean, giving a negative SMB. The net SMB associated with ice formation (by conversion from snow) and melting/runoff is computed for each column, averaged over the coupling interval, and sent to the coupler (*qice, mm/s*). If *glc_dyntopo* is false, then surface runoff for glacier_mec land units is computed as for glacier land units: Any snow in excess of 1 m LWE runs off to the ocean, and Melted ice remains in place until it refreezes. Excess snow and melted ice still contribute to positive and negative values, respectively, of *qice*, but only for the purpose of forcing Glimmer-CISM. Currently, *glc_dyntopo* = false is the default, and the only supported option.

Note that the SMB typically is defined as the total accumulation of ice and snow, minus the total ablation. The *qice* flux passed to Glimmer-CISM is the mass balance for ice alone, not snow. We can think of CLM as owning the

snow, whereas Glimmer-CISM owns the underlying ice. Fluctuations in snow depth between 0 and 1 m LWE are not reflected in the SMB passed to Glimmer-CISM.

2.14 Model for Scale Adaptive River Transport (MOSART)

2.14.1 Overview

MOSART is a river transport model designed for applications across local, regional and global scales ([Li et al., 2013b](#)). A major purpose of MOSART is to provide freshwater input for the ocean model in coupled Earth system model. MOSART also provides an effective way of evaluating and diagnosing the soil hydrology simulated by land surface models through direction comparison of the simulated river flow with observations of natural streamflow at gauging stations ([Li et al., 2015a](#)). Moreover, MOSART provides a modeling framework for representing riverine transport and transformation of energy and biogeochemical fluxes under both natural and human-influenced conditions ([Li et al., 2015b](#)).

2.14.2 Routing Processes

MOSART divides each spatial unit such as a lat/lon grid or watershed into three categories of hydrologic units (as shown in [Figure 2.14.2](#)): hillslopes that contribute both surface and subsurface runoff into tributaries, tributaries that discharge into a single main channel, and the main channel connects the local spatial unit with upstream/downstream units through the river network. MOSART assumes that all the tributaries within a spatial unit can be treated as a single hypothetical sub-network channel with a transport capacity equivalent to all the tributaries combined. Correspondingly, three routing processes are represented in MOSART: 1) hillslope routing: in each spatial unit, surface runoff is routed as overland flow into the sub-network channel, while subsurface runoff generated in the spatial unit directly enters the sub-network channel; 2) sub-network channel routing: the sub-network channel receives water from the hillslopes, routes water through the channel and discharges it into the main channel; 3) main channel routing: the main channel receives water from the sub-network channel and/or inflow, if any, from the upstream spatial units, and discharges the water to its downstream spatial unit or the ocean.

MOSART only route positive runoff, although negative runoff could be generated occasionally by the land model (e.g., q_{gwl}). Negative runoff in any runoff component including q_{sur} , q_{sub} , q_{gwl} is not routed through MOSART, but instead is mapped directly from the spatial unit where it is generated at any time step to the coupler.

In MOSART, the travel velocities of water across hillslopes, sub-network and main channel are all estimated using the Manning's equation with different levels of simplifications. Generally the Manning's equation is in the form of

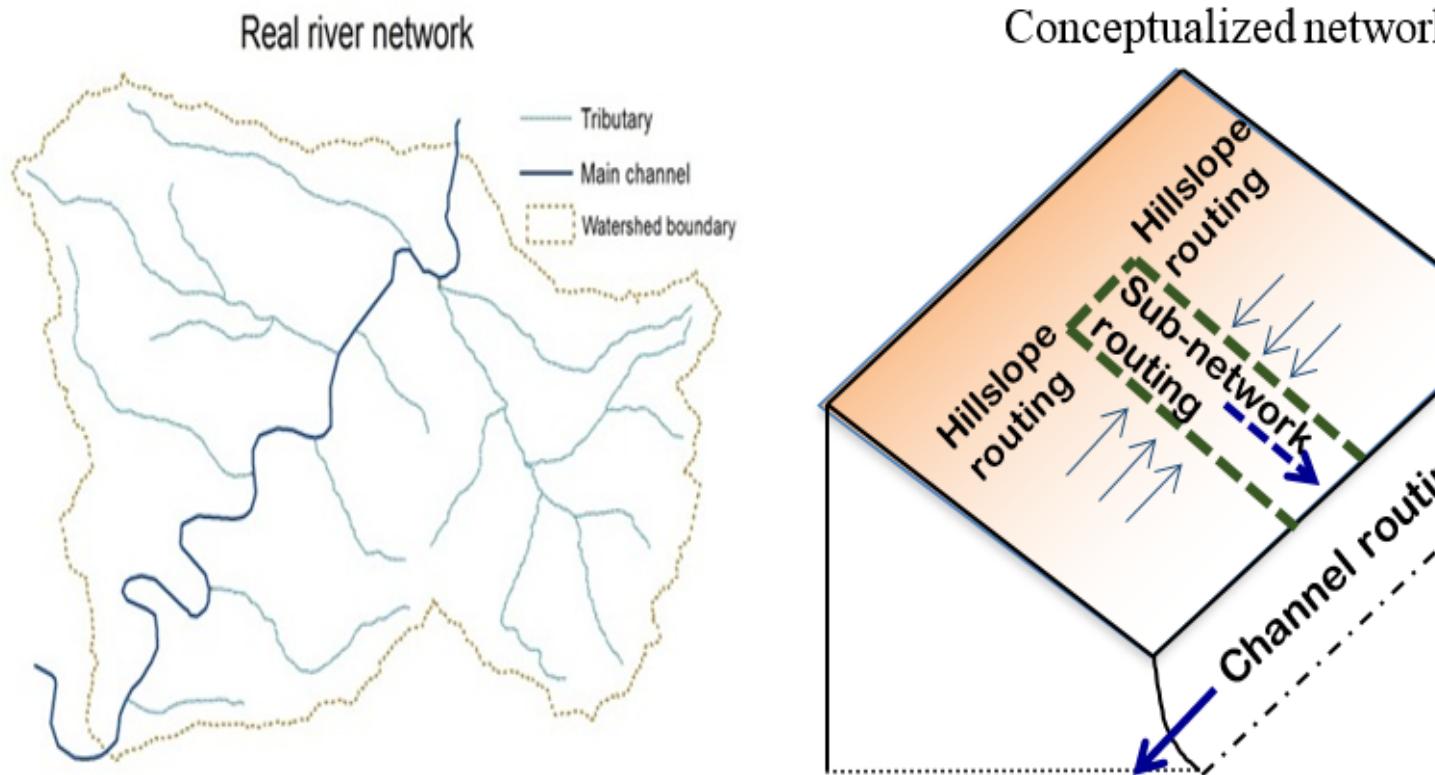
$$V = \frac{R^{\frac{2}{3}} S_f}{n} \quad (2.675)$$

where :math: V is the travel velocity (m s^{-1}), R is the hydraulic radius (m). S_f is the friction slope, and accounting for the effects of gravity, friction, inertia and other forces on the water. If the channel slope is steep enough, the gravity force dominates over the others so one can approximate S_f by the channel bed slope S , which is the key assumption underpinning the kinematic wave method. n is the Manning's roughness coefficient, which is mainly controlled by surface roughness and sinuosity of the flow path.

If the water surface is sufficiently large or the water depth h is sufficiently shallow, the hydraulic radius can be approximated by the water depth. This is the case for both hillslope and sub-network channel routing.

$$R_h = h_h R_t = h_t \quad (2.676)$$

Here R_h (m) and R_t (m) are hydraulic radius for hillslope and sub-network channel routing respectively, and h_h (m) and h_t (m) are water depth during hillslope and sub-network channel routing respectively.



For the main channel, the hydraulic radius is given by

$$R_r = \frac{A_r}{P_r} \quad (2.677)$$

where A_r (m^2) is the wetted area defined as the part of the channel cross-section area below the water surface, P_r (m) is the wetted perimeter (m), the perimeter confines in the wetted area.

For hillslopes, sub-network and main channels, a common continuity equation can be written as

$$\frac{dS}{dt} = Q_{in} - Q_{out} + R \quad (2.678)$$

where Q_{in} ($\text{m}^3 \text{s}^{-1}$) is the main channel flow from the upstream grid(s) into the main channel of the current grid, which is zero for hillslope and sub-network routing. Q_{out} ($\text{m}^3 \text{s}^{-1}$) is the outflow rate from hillslope into the sub-network, from the sub-network into the main channel, or from the current main channel to the main channel of its downstream grid (if not the outlet grid) or ocean (if the current grid is the basin outlet). R ($\text{m}^3 \text{s}^{-1}$) is a source term, which could be the surface runoff generation rate for hillslopes, or lateral inflow (from hillslopes) into sub-network channel or water-atmosphere exchange fluxes such as precipitation and evaporation. It is assumed that surface runoff is generated uniformly across all the hillslopes. Currently, MOSART does not exchange water with the atmosphere or return water to the land model so its function is strictly to transport water from runoff generation through the hillslope, tributaries, and main channels to the basin outlets.

2.14.3 Numerical Solution

The numerical implementation of MOSART is mainly based on a subcycling scheme and a local time-stepping algorithm. There are two levels of subcycling. For convenience, we denote T_{inputs} (s), T_{mosart} (s), $T_{hillslope}$ (s) and

$T_{channel}$ (s) as the time steps of runoff inputs (from CLM to MOSART via the flux coupler), MOSART routing, hillslope routing and channel routing respectively. The first level of subcycling is between the runoff inputs and MOSART routing. If T_{inputs} is 10800s and T_{mosart} is 3600s, three MOSART time steps will be invoked each time the runoff inputs are updated. The second level of subcycling is between the hillslope routing and channel routing. This is to account for the fact that the travel velocity of water across hillslope is usually much slower than that in the channels. $T_{hillslope}$ is usually set as the same as T_{mosart} , but within each time step of hillslope routing there are a few time steps for channel routing, i.e., $T_{hillslope} = D_{levelH2R} \cdot T_{channel}$. The local time-stepping algorithm is to account for the fact that the travel velocity of water is much faster in some river channels (e.g., with steeper bed slope, narrower channel width) than others. That is, for each channel (either a sub-network or main channel), the final time step of local channel routing is given as $T_{local} = T_{channel}/D_{local}$. D_{local} is currently estimated empirically as a function of local channel slope, width, length and upstream drainage area. If MOSART crashes due to a numerical issue, we recommend to increase $D_{levelH2R}$ and, if the issue remains, reducing T_{mosart} .

2.14.4 Parameters and Input Data

MOSART is supported by a comprehensive, global hydrography dataset at 0.5 ° resolution. As such, the fundamental spatial unit of MOSART is a 0.5 ° lat/lon grid. The topographic parameters (such as flow direction, channel length, topographic and channel slopes etc.) were derived using the Dominant River Tracing (DRT) algorithm ([Wu et al., 2011](#) ; [Wu et al. 2012](#)). The DRT algorithm produces the topographic parameters in a scale-consistent way to preserve/upscale the key features of a baseline high-resolution hydrography dataset at multiple coarser spatial resolutions. Here the baseline high-resolution hydrography dataset is the 1km resolution Hydrological data and maps based on SHuttle Elevation Derivatives at multiple Scales (HydroSHEDS) ([Lehner and Döll, 2004](#) ; [Lehner et al., 2008](#)). The channel geometry parameters, e.g., bankfull width and depth, were estimated from empirical hydraulic geometry relationships as functions of the mean annual discharge. The Manning roughness coefficients for overland and channel flow were calculated as functions of landcover and water depth. For more details on the methodology to derive channel geometry and the Manning's roughness coefficients, please refer to [Getirana et al. \(2012\)](#) . The full list of parameters included in this global hydrography dataset is provided in the [Table 2.24](#). Evaluation of global simulations by MOSART using the aforementioned parameters is described in [Li et al. \(2015b\)](#) .

Table 2.24: List of parameters in the global hydrography dataset

Name	Unit	Description
F_{dir}	-	The D8 single flow direction for each coarse grid cell coded using 1 (E), 2 (SE), 4 (S), 8 (SW), 16 (W), 32 (NW), 64 (N), 128 (NE)
A_{total}	km ²	The upstream drainage area of each coarse grid cell
F_{dis}	m	The dominant river length for each coarse grid cell
$S_{channel}$	-	The average channel slope for each coarse grid cell
$S_{topographic}$	-	The average topographic slope (for overland flow routing) for each coarse grid cell
A_{local}	km ²	The surface area for each coarse grid cell
D_p	m ⁻¹	Drainage density, calculated as the total channel length within each coarse grid cell divided by the local cell area
D_r	m	The bankfull depth of main channel
W_r	m	The bankfull width of main channel
D_t	m	The average bankfull depth of tributary channels
W_t	m	The average bankfull width of tributary channels
n_r	-	Manning's roughness coefficient for channel flow routing
n_h	-	Manning's roughness coefficient for overland flow routing

2.14.5 Difference between CLM5.0 and CLM4.5

1. Routing methods: RTM, a linear reservoir method, is used in CLM4.5 for river routing, whilst in CLM5.0, MOSART is an added option for river routing based on the more physically-based kinematic wave method.
2. Runoff treatment: In RTM runoff is routed regardless of its sign so negative streamflow can be simulated at times. MOSART routes only nonnegative runoff and always produces positive streamflow, which is important for future extension for modeling riverine heat and biogeochemical fluxes.
3. Input parameters: RTM in CLM4.5 only requires one layer of spatial variable of channel velocity, whilst MOSART in CLM5.0 requires 13 parameters that are all available globally at 0.5 ° resolution.
4. Outputs: RTM only produces streamflow simulation, whilst MOSART additionally simulates the time-varying channel velocities and channel water depth and channel surface water variation.

2.15 Urban Model (CLMU)

At the global scale, and at the coarse spatial resolution of current climate models, urbanization has negligible impact on climate. However, the urban parameterization (CLMU; [Oleson et al. \(2008b\)](#); [Oleson et al. \(2008c\)](#)) allows simulation of the urban environment within a climate model, and particularly the temperature where people live. As such, the urban model allows scientific study of how climate change affects the urban heat island and possible urban planning and design strategies to mitigate warming (e.g., white roofs).

Urban areas in CLM are represented by up to three urban landunits per gridcell according to density class. The urban landunit is based on the “urban canyon” concept of [Oke \(1987\)](#) in which the canyon geometry is described by building height (H) and street width (W) ([Figure 2.12](#)). The canyon system consists of roofs, walls, and canyon floor. Walls are further divided into shaded and sunlit components. The canyon floor is divided into pervious (e.g., to represent residential lawns, parks) and impervious (e.g., to represent roads, parking lots, sidewalks) fractions. Vegetation is not explicitly modeled for the pervious fraction; instead evaporation is parameterized by a simplified bulk scheme.

Each of the five urban surfaces is treated as a column within the landunit ([Figure 2.12](#)). Radiation parameterizations account for trapping of solar and longwave radiation inside the canyon. Momentum fluxes are determined for the urban landunit using a roughness length and displacement height appropriate for the urban canyon and stability formulations from CLM. A one-dimensional heat conduction equation is solved numerically for a multiple-layer ($N_{levurb} = 10$) column to determine conduction fluxes into and out of canyon surfaces.

A new building energy model has been developed for CLM5.0. It accounts for the conduction of heat through interior surfaces (roof, sunlit and shaded walls, and floors), convection (sensible heat exchange) between interior surfaces and building air, longwave radiation exchange between interior surfaces, and ventilation (natural infiltration and exfiltration). Idealized HAC systems are assumed where the system capacity is infinite and the system supplies the amount of energy needed to keep the indoor air temperature (T_{iB}) within maximum and minimum emperatures ($T_{iB,\max}$, $T_{iB,\min}$), thus explicitly resolving space heating and air conditioning fluxes. Anthropogenic sources of waste heat ($Q_{H,waste}$) from HAC that account for inefficiencies in the heating and air conditioning equipment and from energy lost in the conversion of primary energy sources to end use energy are derived from [Sivak \(2013\)](#). These sources of waste heat are incorporated as modifications to the canyon energy budget.

Turbulent [sensible heat ($Q_{H,u}$) and latent heat ($Q_{E,u}$)] and storage ($Q_{S,u}$) heat fluxes and surface ($T_{u,s}$) and internal ($T_{u,i=1,N_{levgrnd}}$) temperatures are determined for each urban surface u . Hydrology on the roof and canyon floor is simulated and walls are hydrologically inactive. A snowpack can form on the active surfaces. A certain amount of liquid water is allowed to pond on these surfaces which supports evaporation. Water in excess of the maximum ponding depth runs off (R_{roof} , $R_{imprvrd}$, R_{prvrd}).

The heat and moisture fluxes from each surface interact with each other through a bulk air mass that represents air in the urban canopy layer for which specific humidity (q_{ac}) and temperature (T_{ac}) are prognosed ([Figure 2.13](#)). The air temperature can be compared with that from surrounding vegetated/soil (rural) surfaces in the model to ascertain heat island characteristics. As with other landunits, the CLMU is forced either with output from a host atmospheric model (e.g., the Community Atmosphere Model (CAM)) or observed forcing (e.g., reanalysis or field observations). The urban model produces sensible, latent heat, and momentum fluxes, emitted longwave, and reflected solar radiation,

which are area-averaged with fluxes from non-urban “landunits” (e.g., vegetation, lakes) to supply grid cell averaged fluxes to the atmospheric model.

Present day global urban extent and urban properties were developed by [Jackson et al. \(2010\)](#). Urban extent, defined for four classes [tall building district (TBD), and high, medium, and low density (HD, MD, LD)], was derived from LandScan 2004, a population density dataset derived from census data, nighttime lights satellite observations, road proximity, and slope ([Dobson et al. 2000](#)). The urban extent data for TBD, HD, and MD classes are aggregated from the original 1 km resolution to both a 0.05° by 0.05° global grid for high-resolution studies or a 0.5° by 0.5° grid. For the current implementation, the LD class is not used because it is highly rural and better modeled as a vegetated/soil surface. Although the TBD, HD, and MD classes are represented as individual urban landunits, urban model history output is currently a weighted average of the output for individual classes.

For each of 33 distinct regions across the globe, thermal (e.g., heat capacity and thermal conductivity), radiative (e.g., albedo and emissivity) and morphological (e.g., height to width ratio, roof fraction, average building height, and pervious fraction of the canyon floor) properties are provided for each of the density classes. Building interior minimum and maximum temperatures are prescribed based on climate and socioeconomic considerations. The surface dataset creation routines (see CLM5.0 User’s Guide) aggregate the data to the desired resolution.

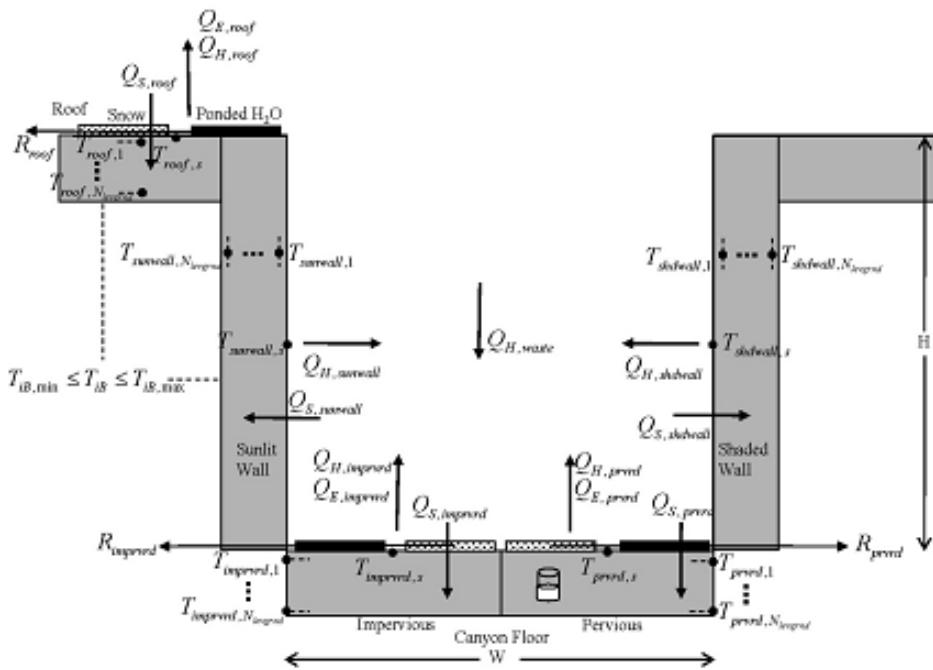


Figure 2.12: Schematic representation of the urban land unit. See the text for description of notation. Incident, reflected, and net solar and longwave radiation are calculated for each individual surface but are not shown for clarity.

The urban model that was first released as a component of CLM4.0 is separately described in the urban technical note ([Oleson et al. \(2010b\)](#)). The main changes in the urban model from CLM4.0 to CLM4.5 were 1) an expansion of the single urban landunit to up to three landunits per grid cell stratified by urban density types, 2) the number of urban layers for roofs and walls was no longer constrained to be equal to the number of ground layers, 3) space heating and

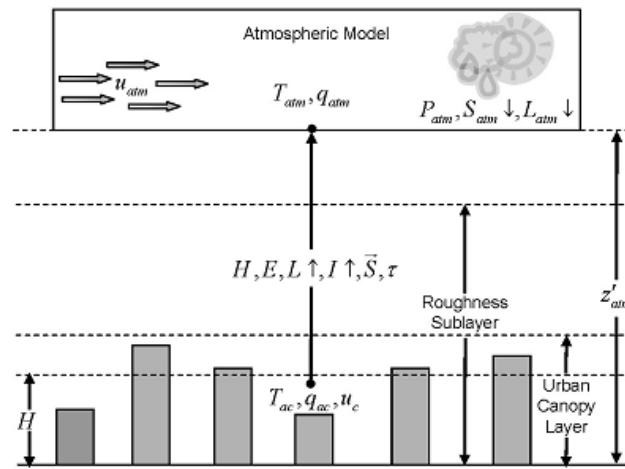


Figure 2.13: Schematic of urban and atmospheric model coupling. The urban model is forced by the atmospheric model wind (u_{atm}), temperature (T_{atm}), specific humidity (q_{atm}), precipitation (P_{atm}), solar ($S_{atm} \downarrow$) and longwave ($L_{atm} \downarrow$) radiation at reference height z'_{atm} (section 2.2.3). Fluxes from the urban landunit to the atmosphere are turbulent sensible (H) and latent heat (λE), momentum (τ), albedo ($I \uparrow$), emitted longwave ($L \uparrow$), and absorbed shortwave (\vec{S}) radiation. Air temperature (T_{ac}), specific humidity (q_{ac}), and wind speed (u_c) within the urban canopy layer are diagnosed by the urban model. H is the average building height.

air conditioning wasteheat factors were set to zero by default so that the user could customize these factors for their own application, 4) the elevation threshold used to eliminate urban areas in the surface dataset creation routines was increased from 2200 meters to 2600 meters, 5) hydrologic and thermal calculations for the pervious road followed CLM4.5 parameterizations.

The main changes in the urban model from CLM4.5 to CLM5.0 are 1) a more sophisticated and realistic building space heating and air conditioning submodel that prognoses interior building air temperature and includes more realistic space heating and air conditioning wasteheat factors (see above), 2) the maximum building temperature (which determines air conditioning demand) is now read in from a namelist-defined file which allows for dynamic control of this input variable. The maximum building temperatures that are defined in [Jackson et al. \(2010\)](#) are implemented in year 1950 (thus air conditioning is off in prior years) and air conditioning is turned off in year 2100 (because the buildings are not suitable for air conditioning in some extreme global warming scenarios). These feature will be described in more detail in a forthcoming paper. In addition, a module of heat stress indices calculated online in the model that can be used to assess human thermal comfort for rural and urban areas has been added. This last development is described and evaluated by [Buzan et al. \(2015\)](#).

2.16 CN Pools

2.16.1 Introduction

CLM includes a prognostic treatment of the terrestrial carbon and nitrogen cycles including natural vegetation, crops, and soil biogeochemistry. The model is fully prognostic with respect to all carbon and nitrogen state variables in the vegetation, litter, and soil organic matter. The seasonal timing of new vegetation growth and litterfall is also prognostic, responding to soil and air temperature, soil water availability, daylength, and crop management practices in varying degrees depending on a specified phenology type or management for each PFT (Chapter 2.20). The prognostic LAI, SAI, tissue stoichiometry, and vegetation heights are utilized by the biophysical model that couples carbon, water, and energy cycles.

Separate state variables for C and N are tracked for leaf, live stem, dead stem, live coarse root, dead coarse root, fine root, and grain pools (Figure 2.14). Each of these pools has two corresponding storage pools representing, respectively, short-term and long-term storage of non-structural carbohydrates and labile nitrogen. There are two additional carbon pools, one for the storage of growth respiration reserves, and another used to meet excess demand for maintenance respiration during periods with low photosynthesis. One additional nitrogen pool tracks retranslocated nitrogen, mobilized from leaf tissue prior to abscission and litterfall. Altogether there are 23 state variables for vegetation carbon, and 22 for vegetation nitrogen.

In addition to the vegetation pools, CLM includes a series of decomposing carbon and nitrogen pools as vegetation successively breaks down to CWD, and/or litter, and subsequently to soil organic matter. Discussion of the decomposition model, alternate specifications of decomposition rates, and methods to rapidly equilibrate the decomposition model, is in Chapter 2.21.

2.16.2 Tissue Stoichiometry

As of CLM5, vegetation tissues have a flexible stoichiometry, as described in [Ghimire et al. \(2016\)](#). Each tissue has a target C:N ratio, and nitrogen is allocated at each timestep in order to allow the plant to best match the target stoichiometry. Nitrogen downregulation of productivity acts by increasing the C:N ratio of leaves when insufficient nitrogen is available to meet stoichiometric demands of leaf growth, thereby reducing the N available for photosynthesis and reducing the $V_{c,max25}$ and J_{max25} terms, as described in Chapter 2.10. Details of the flexible tissue stoichiometry are described in Chapter 2.19.

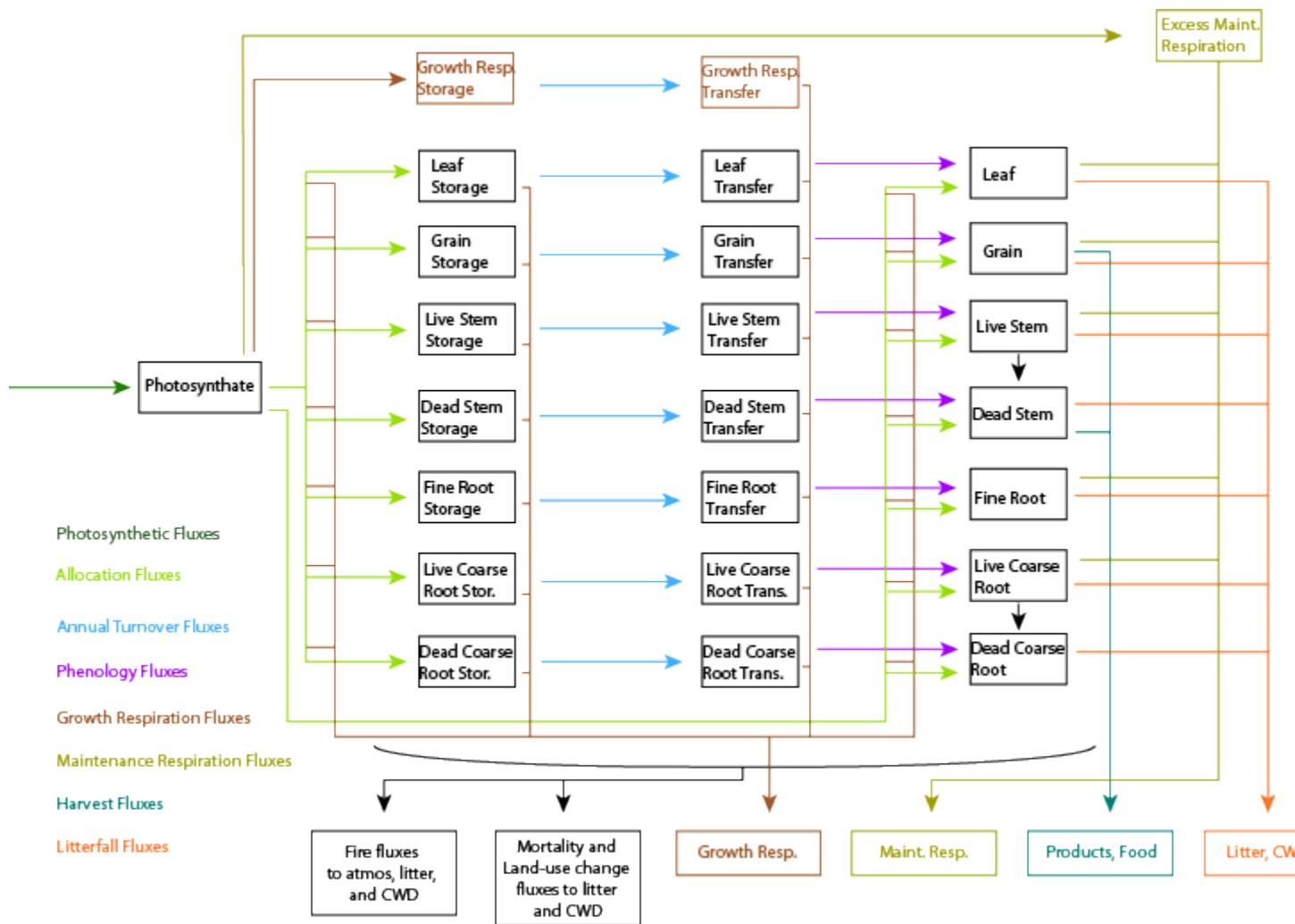


Figure 2.14: Vegetation fluxes and pools for carbon cycle in CLM5.

2.17 Plant Respiration

2.17.1 Autotrophic Respiration

The model treats maintenance and growth respiration fluxes separately, even though it is difficult to measure them as separate fluxes (Lavigne and Ryan, 1997; Sprugel et al., 1995). Maintenance respiration is defined as the carbon cost to support the metabolic activity of existing live tissue, while growth respiration is defined as the additional carbon cost for the synthesis of new growth.

Maintenance Respiration

Atkin et al. (2016) propose a model for respiration that is based on the leaf nitrogen content per unit area (NS_{narea} (gN m⁻² leaf)), with an intercept parameter that is PFT dependant, and an acclimation term that depends upon the average temperature of the previous 10 day period $t_{2m,10days}$, in Celsius.

$$CF_{mr_leaf} = i_{atkin,pft} + (NS_{narea} \cdot 0.2061) - (0.0402(t_{2m,10days})) \quad (2.679)$$

The temperature dependance of leaf maintenance (dark) respiration is described in Chapter 2.9.

$$CF_{mr_livestem_} = NS_{livestem} MR_{base} MR_{Q10}^{(T_{2m}-20)/10} \quad (2.680)$$

$$CF_{mr_livecroo_} = NS_{livecroo} MR_{base} MR_{Q10}^{(T_{2m}-20)/10} \quad (2.681)$$

$$CF_{mr_froot_} = \sum_{j=1}^{nlevsoi} NS_{frootrootfr_j} MR_{base} MR_{Q10}^{(Ts_j-20)/10} \quad (2.682)$$

where MR_{q10} (= 2.0) is the temperature sensitivity for maintenance respiration, T_{2m} (°C) is the air temperature at 2m height, Ts_j (: sup : 'oC) is the soil temperature at level j , and $rootfr_j$ is the fraction of fine roots distributed in soil level j .

Table 2.25: Atkin leaf respiration model intercept values.

Plant functional type	i_{atkin}
NET Temperate	1.499
NET Boreal	1.499
NDT Boreal	1.499
BET Tropical	1.756
BET temperate	1.756
BDT tropical	1.756
BDT temperate	1.756
BDT boreal	1.756
BES temperate	2.075
BDS temperate	2.075
BDS boreal	2.075
C ₃ arctic grass	2.196
C ₃ grass	2.196
C ₄ grass	2.196

Note that, for woody vegetation, maintenance respiration costs are not calculated for the dead stem and dead coarse root components. These components are assumed to consist of dead xylem cells, with no metabolic function. By

separating the small live component of the woody tissue (ray parenchyma, phloem, and sheathing lateral meristem cells) from the larger fraction of dead woody tissue, it is reasonable to assume a common base maintenance respiration rate for all live tissue types.

The total maintenance respiration cost is then given as:

$$CF_{mr} = CF_{mr_leaf} + CF_{mr_froot} + CF_{mr_livestem} + CF_{mr_livecroot}. \quad (2.683)$$

Growth Respiration

Growth respiration is calculated as a factor of 0.11 times the total carbon allocation to new growth (CF_{growth} , after allocating carbon for N acquisition, Chapter 2.18.) on a given timestep, based on construction costs for a range of woody and non-woody tissues (Atkin et al. in prep). For new carbon and nitrogen allocation that enters storage pools for subsequent display, it is not clear what fraction of the associated growth respiration should occur at the time of initial allocation, and what fraction should occur later, at the time of display of new growth from storage. Eddy covariance estimates of carbon fluxes in forest ecosystems suggest that the growth respiration associated with transfer of allocated carbon and nitrogen from storage into displayed tissue is not significant (Churkina et al., 2003), and so it is assumed in CLM that all of the growth respiration cost is incurred at the time of initial allocation, regardless of the fraction of allocation that is displayed immediately (i.e. regardless of the value of f_{cur} , section 13.5). This behavior is parameterized in such a way that if future research suggests that some fraction of the growth respiration cost should be incurred at the time of display from storage, a simple parameter modification will effect the change.¹

2.18 Fixation and Uptake of Nitrogen (FUN)

2.18.1 Introduction

The Fixation and Uptake of Nitrogen model is based on work by [Fisher et al. \(2010\)](#), [Brzostek et al. \(2014\)](#), and [Shi et al. \(2016\)](#). The concept of FUN is that in most cases, Nitrogen uptake requires the expenditure of energy in the form of carbon, and further, that there are numerous potential sources of Nitrogen in the environment which a plant may exchange for carbon. The ratio of carbon expended to Nitrogen acquired is referred to here as the cost, or exchange rate, of N acquisition (E_{nacq} , gC/gN)). There are eight pathways for N uptake:

1. Fixation by symbiotic bacteria in root nodules (for N fixing plants) ($_{fix}$)
2. Retranslocation of N from senescing tissues ($_{ret}$)
3. Active uptake of NH₄ by arbuscular mycorrhizal plants ($_{active,nh4}$)
4. Active uptake of NH₄ by ectomycorrhizal plants ($_{active,nh4}$)
5. Active uptake of NO₃ by arbuscular mycorrhizal plants ($_{active,no3}$)
6. Active uptake of NO₃ by ectomycorrhizal plants ($_{active,no3}$)
7. Nonmycorrhizal uptake of NH₄ ($_{nonmyc,no3}$)
8. Nonmycorrhizal uptake of NO₃ ($_{nonmyc,no3}$)

The notation suffix for each pathway is given in parentheses here. At each timestep, each of these pathways is associated with a cost term ($N_{cost,x}$), a payment in carbon ($C_{nuptake,x}$), and an influx of Nitrogen ($N_{uptake,x}$) where x is one of the eight uptake streams listed above.

For each PFT, we define a fraction of the total C acquisition that can be used for N fixation (f_{fixers}), which is broadly equivalent to the fraction of a given PFT that is capable of fixing Nitrogen, and thus represents an upper limit on the

¹ Parameter grpnow in routines CNGResp and CNAllocation, currently set to 1.0, could be changed to a smaller value to transfer some portion (1 - grpnow) of the growth respiration forward in time to occur at the time of growth display from storage.

amount to which fixation can be increased in low n conditions. For each PFT, the cost calculation is conducted twice. Once where fixation is possible and once where it is not. (f_{fixers})

For all of the active uptake pathways, whose cost depends on varying concentrations of N through the soil profile, the costs and fluxes are also determined by soil layer j .

2.18.2 Boundary conditions of FUN

Available Carbon

The carbon available for FUN, C_{avail} (gC m⁻²) is the total canopy photosynthetic uptake (GPP), minus the maintenance respiration fluxes (m_r) and multiplied by the time step in seconds (δt). Thus, the remainder of this chapter considers fluxes per timestep, and integrates these fluxes as they are calculated.

$$C_{avail} = (GPP - m_r)\delta t$$

Growth respiration is thus only calculated on the part of the carbon uptake that remains after expenditure of C by the FUN module.

Available Soil Nitrogen

Cost of Nitrogen Fixation

The cost of fixation is derived from *Houlton et al. (2008)*.

$$N_{cost,fix} = -s_{fix}/(1.25e^{a_{fix}+b_{fix}.t_{soil}(1-0.5t_{soil}/c_{fix})})$$

Herein, a_{fix} , b_{fix} and c_{fix} are all parameters of the temperature response function of fixation reported by Houlton et al. (2008) ($\exp[a + bT_s(1 - 0.5T_s/c)]$). $t_{\{soil\}}$ is the soil temperature in C. The values of these parameters are fitted to empirical data as $a=-3.62 \pm 0.52$, $b=0.27$ and $c=25.15 \pm 0.66$. 1.25 converts from the temperature response function to a 0-1 limitation factor (as specifically employed by Houlton et al.). This function is a ‘rate’ of uptake for a given temperature. Here we assimilated the rate of fixation into the cost term by assuming that the rate is analogous to a conductance for N, and inverting the term to produce a cost/resistance analogue. We then multiply this temperature term by the minimum cost at optimal temperature (s_{fix}) to give a temperature limited cost in terms of C to N ratios.

Cost of Active Uptake

The cost of N uptake from soil, for each layer j , is controlled by two uptake parameters that pertain respectively to the relationship between soil N content and N uptake, and root C density and N uptake.

For non-mycorrhizal uptake:

$$N_{cost,nonmyc,j} = \frac{k_{n,nonmyc}}{N_{smin,j}} + \frac{k_{c,nonmyc}}{c_{root,j}}$$

and for active uptake:

$$N_{cost,active,j} = \frac{k_{n,active}}{N_{smin,j}} + \frac{k_{c,active}}{c_{root,j}}$$

where $k_{n,active}$ varies according to whether we are considering ecto or arbuscular mycorrhizal uptake.

$$k_{n,active} = \begin{cases} k_{n,Eactive} & e = 1 \\ k_{n,Aactive} & e = 0 \end{cases} \quad (2.684)$$

where m=1 pertains to the fraction of the PFT that is ecotmycorrhizal, as opposed to arbuscular mycorrhizal.

2.18.3 Resolving N cost across simultaneous uptake streams

The total cost of N uptake is calculated based on the assumption that carbon is partitioned to each stream in proportion to the inverse of the cost of uptake. So, more expensive pathways receive less carbon. Earlier versions of FUN (Fisher et al., 2010)<Fisheral2010>) utilized a scheme whereby plants only took up N from the cheapest pathway. Brzostek et al. (2014) introduced a scheme for the simultaneous uptake from different pathways. Here we calculate a ‘conductance’ to N uptake (analogous to the inverse of the cost function conceptualized as a resistance term) $N_{conductance}$ (gN/gC) as:

$$N_{conductance,f} = \sum (1/N_{cost,x})$$

From this, we then calculate the fraction of the carbon allocated to each pathway as

$$C_{frac,x} = \frac{1/N_{cost,x}}{N_{conductance}}$$

These fractions are used later, to calculate the carbon expended on different uptake pathways. Next, the N acquired from each uptake stream per unit C spent ($N_{exch,x}$, gN/gC) is determined as

$$N_{exch,x} = \frac{C_{frac,x}}{N_{cost,x}}$$

We then determine the total amount of N uptake per unit C spent ($N_{exch,tot}$, gN/gC) as the sum of all the uptake streams.

$$N_{exch,tot} = \sum N_{exch,x}$$

and thus the subsequent overall N cost is

$$N_{cost,tot} = 1/N_{exch,tot}$$

Retranslocation is determined via a different set of mechanisms, once the $N_{cost,tot}$ is known.

2.18.4 Nitrogen Retranslocation

The retranslocation uses an iterative algorithm to remove Nitrogen from each piece of falling litter. There are two pathways for this, ‘free’ uptake which removes the labile N pool, and ‘paid-for’ uptake which uses C to extract N from increasingly more recalcitrant pools.

At each timestep, the pool of carbon in falling leaves ($C_{fallingleaf}$, g m⁻²) is generated from the quantity of litterfall on that day (see Phenology chapter for details). The amount of N in the litter pool ($N_{fallingleaf}$, g m⁻²) is calculated as the total leaf N multiplied by the fraction of the leaf pool passed to litter that timestep.

$$N_{fallingleaf} = N_{leaf} \cdot C_{fallingleaf} / C_{leaf}$$

The carbon available at the beginning of the iterative retranslocation calculation is equal to the C_{avail} input into FUN.

$$C_{avail,retrans,0} = C_{avail}$$

Free Retranslocation

Some part of the leaf Nitrogen pool is removed without the need for an C expenditure. This ‘free’ N uptake amount, ($N_{retrans,free}$, gN m⁻²) is calculated as

$$N_{retrans,free} = \max(N_{fallingleaf} - (C_{fallingleaf} / CN_{litter,min}), 0.0)$$

where $CN_{litter,min}$ is the minimum C:N ratio of the falling litter (currently set to 1.5 x the target C:N ratio).

The new $N_{fallingleaf}$ (gN m⁻²) is then determined as

$$N_{fallingleaf} = N_{fallingleaf} - N_{retrans,free}$$

and the new litter C:N ratio as

$$CN_{fallingleaf} = C_{fallingleaf} / N_{fallingleaf}$$

Paid-for Retranslocation

The remaining calculations conduct an iterative calculation to determine the degree to which N retranslocation from leaves is paid for as C:N ratios and thus cost increase as N is extracted. The iteration continues until either

1. The cost of retranslocation ($cost_{retrans}$ increases beyond the cost of acquiring N from alternative pathways ($N_{cost,tot}$)).
2. $CN_{fallingleaf}$ rises to a maximum level, after which no more extraction is possible (representing unavoidable N loss) or
3. There is no more carbon left to pay for extraction.

First we calculate the cost of extraction ($cost_{retrans}$, gC/gN) for the current leaf C:N ratio as

$$cost_{retrans} = k_{retrans}/(1/CN_{fallingleaf})^{1.3}$$

where $k_{retrans}$ is a parameter controlling the overall cost of resorption, which also increases exponentially as the C:N ratio increases **Say something about 1.3 exponent**.

Next, we calculate the amount of C needed to be spent to increase the falling leaf C:N ratio by 1.0 in this iteration i ($C_{retrans,spent,i}$)

$$C_{retrans,spent,i} = cost_{retrans} \cdot (N_{fallingleaf} - C_{fallingleaf}/(CN_{fallingleaf} + 1.0))$$

(wherein the retranslocation cost is assumed to not change over the increment of 1.0 in C:N ratio). Next, we calculate whether this is larger than the remaining C available to spend.

$$C_{retrans,spent,i} = \min(C_{retrans,spent,i}, C_{avail,retrans,i})$$

The amount of N retranslocated from the leaf in this iteration ($N_{retrans,paid,i}$, gN m⁻²) is calculated, checking that it does not fall below zero:

$$N_{retrans,paid,i} = \min(N_{fallingleaf}, C_{retrans,spent,i}/cost_{retrans})$$

The next step calculates the growth C which is accounted for by this amount of N extraction in this iteration ($C_{retrans,accounted,i}$). This is calculated using the current plant C:N ratio, and also for the additional C which will need to be spent on growth respiration to build this amount of new tissue.

$$C_{retrans,accounted,i} = N_{retrans,paid,i} \cdot CN_{plant} \cdot (1.0 + gr_{frac})$$

Then the falling leaf N is updated:

$$N_{fallingleaf} = N_{fallingleaf} - N_{ret,i}$$

and the $CN_{fallingleaf}$ and $cost_{retrans}$ are updated. The amount of available carbon that is either unspent on N acquisition nor accounted for by N uptake is updated:

$$C_{avail,retrans,i+1} = C_{avail,retrans,i} - C_{retrans,spent,i} - C_{retrans,accounted,i}$$

Outputs of Retranslocation algorithm.

The final output of the retranslocation calculation are the retranslocated N ($N_{retrans}$, gN m⁻²), C spent on retranslocation ($C_{retrans,paid}$, gC m⁻²), and C accounted for by retranslocation ($C_{retrans,accounted}$, gC m⁻²).

For paid-for uptake, we accumulate the total carbon spent on retranslocation ($C_{spent,retrans}$),

$$C_{retrans,spent} = \sum C_{retrans,i}$$

The total N acquired from retranslocation is

$$N_{retrans} = N_{retrans,paid} + N_{retrans,free}$$

where N acquired by paid-for retranslocation is

$$N_{retrans,paid} = \sum N_{retrans,paid,i}$$

The total carbon accounted for by retranslocation is the sum of the C accounted for by paid-for N uptake ($N_{retrans,paid}$) and by free N uptake ($N_{retrans,free}$).

$$C_{retrans,accounted} = \sum C_{retrans,accounted,i} + N_{retrans,free} \cdot CN_{plant} \cdot (1.0 + grfrac)$$

The total available carbon in FUN to spend on fixation and active uptake ($C_{tospend}$, gC m⁻²) is calculated as the carbon available minus that account for by retranslocation:

$$C_{tospend} = C_{avail} - C_{retrans,accounted}$$

2.18.5 Carbon expenditure on fixation and active uptake.

At each model timestep, the overall cost of N uptake is calculated (see below) in terms of C:N ratios. The available carbon (C_{avail} , g m⁻² s⁻¹) is then allocated to two alternative outcomes, payment for N uptake, or conservation for growth. For each carbon conserved for growth, a corresponding quantity of N must be made available. In the case where the plant target C:N ratio is fixed, the partitioning between carbon for growth (C_{growth}) and carbon for N uptake ($C_{nuptake}$) is calculated by solving a system of simultaneous equations. First, the carbon available must equal the carbon spent on N uptake plus that saved for growth.

$$C_{growth} + C_{nuptake} = C_{avail}$$

Second, the nitrogen acquired from expenditure of N (left hand side of term below) must equal the N that is required to match the growth carbon (right hand side of term below).

$$C_{nuptake}/N_{cost} = C_{growth}/CN_{target}$$

The solution to these two equated terms can be used to estimate the ideal $C_{nuptake}$ as follows,

$$C_{nuptake} = C_{tospend} / ((1.0 + f_{gr} * (CN_{target}/N_{cost})) + 1).$$

and the other C and N fluxes can be determined following the logic above.

2.18.6 Modifications to allow variation in C:N ratios

The original FUN model as developed by [Fisher et al. \(2010\)](#) and [Brzostek et al. \(2014\)](#) assumes a fixed plant tissue C:N ratio. This means that in the case where N is especially limiting, all excess carbon will be utilized in an attempt to take up more Nitrogen. It has been repeatedly observed, however, that in these circumstances in real life, plants have some flexibility in the C:N stoichiometry of their tissues, and therefore, this assumption may not be realistic. [lit review on CN ratios](#)

Thus, in CLM5, we introduce the capacity for tissue C:N ratios to be prognostic, rather than static. Overall N and C availability (N_{uptake} and C_{growth}) and hence tissue C:N ratios, are both determined by FUN. Allocation to individual tissues is discussed in the allocation chapter

Here we introduce an algorithm which adjusts the C expenditure on uptake to allow varying tissue C:N ratios. Increasing C spent on uptake will directly reduce the C:N ratio, and reducing C spent on uptake (retaining more for tissue growth) will increase it. C spent on uptake is impacted by both the N cost in the environment, and the existing tissue C:N ratio of the plant. The output of this algorithm is γ_{FUN} , the fraction of the ideal $C_{nuptake}$ calculated from the FUN equation above ([link equation](#)).

$$C_{nuptake} = C_{nuptake} \cdot \gamma_{FUN}$$

Response of C expenditure to Nitrogen uptake cost

The environmental cost of Nitrogen ($N_{cost,tot}$) is used to determine γ_{FUN} .

$$\gamma_{FUN} = \max(0.0, 1.0 - (N_{cost,tot} - a_{cnflex})/b_{cnflex})$$

where a_{cnflex} and b_{cnflex} are parameters fitted to give flexible C:N ranges over the operating range of N costs of the model. Calibration of these parameters should be subject to future testing in idealized experimental settings; they are here intended as a placeholder to allow some flexible stoichiometry, in the absence of adequate understanding of this process. Here a_{cnflex} operates as the $N_{cost,tot}$ above which there is a modification in the C expenditure (to allow higher C:N ratios), and b_{cnflex} is the scalar which determines how much the C expenditure is modified for a given discrepancy between a_{cnflex} and the actual cost of uptake.

Response of C expenditure to plant C:N ratios

We first calculate a δ_{CN} , which is the difference between the target C:N ($target_{CN}$) a model parameter, and the existing C:N ratio (CN_{plant}) **This isn't strictly how it is worked out. Need to remember why we use c_allometry instead.**

$$CN_{plant} = \frac{C_{leaf} + C_{leaf,storage}}{N_{leaf} + N_{leaf,storage}}$$

and

$$\delta_{CN} = CN_{plant} - target_{CN}$$

We then increase γ_{FUN} to account for situations where (even if N is expensive) plant C:N ratios have increased too far from the target. Where δ_{CN} is negative, we reduce C spent on N uptake and retain more C for growth

$$\gamma_{FUN} = \begin{cases} \gamma_{FUN} + 0.5.(delta_{CN}/c_{flexcn}) & delta_{CN} > 0 \\ \gamma_{FUN} + (1 - \gamma_{FUN}).min(1, \delta_{CN}/c_{flexcn}) & delta_{CN} < 0 \end{cases}$$

We then restrict the degree to which C expenditure can be reduced (to prevent unrealistically high C:N ratios) as

$$\gamma_{FUN} = max(min(1.0, \gamma_{FUN}), 0.5)$$

2.18.7 Calculation of N uptake streams from active uptake and fixation

Once the final $C_{nuptake}$ is known, the fluxes of C to the individual pools can be derived as

$$C_{nuptake,x} = C_{frac,x} \cdot C_{nuptake}$$

$$N_{uptake,x} = \frac{C_{nuptake}}{N_{cost}}$$

Following this, we determine whether the extraction estimates exceed the pool size for each source of N. Where $N_{active,no3} + N_{nonmyc,no3} > N_{avail,no3}$, we calculate the unmet uptake, $N_{unmet,no3}$

$$N_{unmet,no3} = N_{active,no3} + N_{nonmyc,no3} - N_{avail,no3}$$

then modify both fluxes to account

$$N_{active,no3} = N_{active,no3} + N_{unmet,no3} \cdot \frac{N_{active,no3}}{N_{active,no3} + N_{nonmyc,no3}}$$

$$N_{nonmyc,no3} = N_{nonmyc,no3} + N_{unmet,no3} \cdot \frac{N_{nonmyc,no3}}{N_{active,no3} + N_{nonmyc,no3}}$$

and similarly, for NH4, where $N_{active,nh4} + N_{nonmyc,nh4} > N_{avail,nh4}$, we calculate the unmet uptake, $N_{unmet,no3}$

$$N_{unmet,nh4} = N_{active,nh4} + N_{nonmyc,nh4} - N_{avail,nh4}$$

then modify both fluxes to account

$$N_{active,nh4} = N_{active,nh4} + N_{unmet,nh4} \cdot \frac{N_{active,nh4}}{N_{active,nh4} + N_{nonmyc,nh4}}$$

$$N_{nonmyc,nh4} = N_{nonmyc,nh4} + N_{unmet,nh4} \cdot \frac{N_{nonmyc,nh4}}{N_{active,nh4} + N_{nonmyc,nh4}}$$

and then update the C spent to account for the new lower N acquisition in that layer/pool.

$$\begin{aligned} C_{active,nh4} &= N_{active,nh4} \cdot N_{cost,active,nh4} \\ C_{active,no3} &= N_{active,no3} \cdot N_{cost,active,no3} \\ C_{nonmyc,no3} &= N_{nonmyc,no3} \cdot N_{cost,nonmyc,no3} \\ C_{nonmyc,no3} &= N_{nonmyc,no3} \cdot N_{cost,nonmyc,no3} \end{aligned}$$

Following this, we determine how much carbon is accounted for for each soil layer.

$$C_{accounted,x,j} = C_{spent,j,x} - (N_{acquired,j,x} \cdot CN_{plant} \cdot (1.0 + grfrac))$$

2.18.8 Types of N uptake streams

Arbuscular mycorrhizal fungi: Ectomycorrhizal fungi: Nonmycorrhizal plants.

ECK_active (step 1) sets active components for Ectomycorrhizal fungi ACK_active (step 2) sets active components for Arbuscular fungi

kc_nonmyc (step 1) sets nonmyc components for Ectomycorrhizal fungi kc_nonmyc (step 2) sets active components for Arbuscular fungi

ACTIVE vs NONMYC ECTO vs ARBU for ACTIVE.

2.19 Carbon and Nitrogen Allocation

2.19.1 Introduction

The carbon and nitrogen allocation routines in CLM determine the fate of newly assimilated carbon, coming from the calculation of photosynthesis, and available mineral nitrogen, coming from plant uptake of mineral nitrogen in the soil or being drawn out of plant reserves. A significant change to CLM5 relative to prior versions is that allocation of carbon and nitrogen proceed independently rather than in a sequential manner.

2.19.2 Carbon Allocation for Maintenance Respiration Costs

Allocation of available carbon on each time step is prioritized, with first priority given to the demand for carbon to support maintenance respiration of live tissues (section 13.7). Second priority is to replenish the internal plant carbon pool that supports maintenance respiration during times when maintenance respiration exceeds photosynthesis (e.g. at night, during winter for perennial vegetation, or during periods of drought stress) (Sprugel et al., 1995). Third priority is to support growth of new tissues, including allocation to storage pools from which new growth will be displayed in subsequent time steps.

The total maintenance respiration demand (CF_{mr} , gC m⁻² s⁻¹) is calculated as a function of tissue mass and nitrogen concentration, and temperature (section 13.7). The carbon supply to support this demand is composed of fluxes allocated from carbon assimilated in the current timestep ($CF_{GPP,mr}$, gC m⁻² s⁻¹) and from a storage pool that is drawn down when total demand exceeds photosynthesis ($CF_{xs,mr}$, gC m⁻² s⁻¹):

$$CF_{mr} = CF_{GPP,mr} + CF_{xs,mr} \quad (2.685)$$

$$CF_{GPP,mr} = \begin{cases} CF_{mr} & \text{for } CF_{mr} \leq CF_{GPP} \\ CF_{GPP} & \text{for } CF_{mr} > CF_{GPP} \end{cases} \quad (2.686)$$

$$CF_{xs,mr} = - \begin{cases} 0 & \text{for } CF_{mr} \leq CF_{GPP} \\ CF_{mr} - CF_{GPP} & \text{for } CF_{mr} > CF_{GPP} \end{cases} \quad (2.687)$$

The storage pool that supplies carbon for maintenance respiration in excess of current CF_{GPP} (CS_{xs} , gC m⁻²) is permitted to run a deficit (negative state), and the magnitude of this deficit determines an allocation demand which gradually replenishes CS_{xs} . The logic for allowing a negative state for this pool is to eliminate the need to know in advance what the total maintenance respiration demand will be for a particular combination of climate and plant type. Using the deficit approach, the allocation to alleviate the deficit increases as the deficit increases, until the supply of carbon into the pool balances the demand for carbon leaving the pool in a quasi-steady state, with variability driven by the seasonal cycle, climate variation, disturbance, and internal dynamics of the plant-litter-soil system. In cases where the combination of climate and plant type are not suitable to sustained growth, the deficit in this pool increases until the available carbon is being allocated mostly to alleviate the deficit, and new growth approaches zero. The allocation flux to CS_{xs} ($CF_{GPP,xs}$, gC m⁻² s⁻¹) is given as

$$CF_{GPP,xs,pot} = \begin{cases} 0 & \text{for } CS_{xs} \geq 0 \\ -CS_{xs}/(86400\tau_{xs}) & \text{for } CS_{xs} < 0 \end{cases} \quad (2.688)$$

$$CF_{GPP,xs} = \begin{cases} CF_{GPP,xs,pot} & \text{for } CF_{GPP,xs,pot} \leq CF_{GPP} - CF_{GPP,mr} \\ \max(CF_{GPP} - CF_{GPP,mr}, 0) & \text{for } CF_{GPP,xs,pot} > CF_{GPP} - CF_{GPP,mr} \end{cases} \quad (2.689)$$

where τ_{xs} is the time constant (currently set to 30 days) controlling the rate of replenishment of CS_{xs} .

Note that these two top-priority carbon allocation fluxes ($CF_{GPP,mr}$ and $CF_{GPP,xs}$) are not stoichiometrically associated with any nitrogen fluxes.

2.19.3 Carbon and Nitrogen Stoichiometry of New Growth

After accounting for the carbon cost of maintenance respiration, the remaining carbon flux from photosynthesis which can be allocated to new growth (CF_{avail} , gC m⁻² s⁻¹) is

$$CF_{avail_alloc} = CF_{GPP} - CF_{GPP,mr} - CF_{GPP,xs}. \quad (2.690)$$

Potential allocation to new growth is calculated for all of the plant carbon and nitrogen state variables based on specified C:N ratios for each tissue type and allometric parameters that relate allocation between various tissue types. The allometric parameters are defined as follows:

$$\begin{aligned} a_1 &= \text{ratio of new fine root : new leaf carbon allocation} \\ a_2 &= \text{ratio of new coarse root : new stem carbon allocation} \\ a_3 &= \text{ratio of new stem : new leaf carbon allocation} \\ a_4 &= \text{ratio new live wood : new total wood allocation} \\ g_1 &= \text{ratio of growth respiration carbon : new growth carbon.} \end{aligned} \quad (2.691)$$

Parameters a_1 , a_2 , and a_4 are defined as constants for a given PFT (Table 13.1), while $g_1 = 0.3$ (unitless) is prescribed as a constant for all PFTs, based on construction costs for a range of woody and non-woody tissues (Larcher, 1995).

The model includes a dynamic allocation scheme for woody vegetation (parameter $a_3 = -1$, Table 2.26), in which case the ratio for carbon allocation between new stem and new leaf increases with increasing net primary production (NPP), as

$$a_3 = \frac{2.7}{1 + e^{-0.004NPP_{ann}-300}} - 0.4 \quad (2.692)$$

where NPP_{ann} is the annual sum of NPP from the previous year. This mechanism has the effect of increasing woody allocation in favorable growth environments (Allen et al., 2005; Vanninen and Makela, 2005) and during the phase of stand growth prior to canopy closure (Axelsson and Axelsson, 1986).

Table 2.26: Allocation and target carbon:nitrogen ratio parameters

Plant functional type	a_1	a_2	a_3	a_4	$TargetCN_{leaf}$	$TargetCN_{fr}$	$TargetCN_{lw}$	$TargetCN_{dw}$
NET Temperate	1	0.3	-1	0.1	35	42	50	500
NET Boreal	1	0.3	-1	0.1	40	42	50	500
NDT Boreal	1	0.3	-1	0.1	25	42	50	500
BET Tropical	1	0.3	-1	0.1	30	42	50	500
BET temperate	1	0.3	-1	0.1	30	42	50	500
BDT tropical	1	0.3	-1	0.1	25	42	50	500
BDT temperate	1	0.3	-1	0.1	25	42	50	500
BDT boreal	1	0.3	-1	0.1	25	42	50	500
BES temperate	1	0.3	0.2	0.5	30	42	50	500
BDS temperate	1	0.3	0.2	0.5	25	42	50	500
BDS boreal C ₃ arctic grass	1 1	0.3 0	0.2 0	0.1 0	25 25	42 42	50 0	500 0
C ₃ grass	2	0	0	0	25	42	0	0
C ₄ grass	2	0	0	0	25	42	0	0
Crop R	2	0	0	0	25	42	0	0
Crop I	2	0	0	0	25	42	0	0
Corn R	2	0	0	1	25	42	50	500
Corn I	2	0	0	1	25	42	50	500
Temp Cereal R	2	0	0	1	25	42	50	500
Temp Cereal I	2	0	0	1	25	42	50	500
Winter Cereal R	2	0	0	1	25	42	50	500
Winter Cereal I	2	0	0	1	25	42	50	500
Soybean R	2	0	0	1	25	42	50	500
Soybean I	2	0	0	1	25	42	50	500

Carbon to nitrogen ratios are defined for different tissue types as follows:

$$\begin{aligned} CN_{leaf} &= \text{C : N for leaf} \\ CN_{fr} &= \text{C : N for fine root} \\ CN_{lw} &= \text{C : N for live wood (in stem and coarse root)} \\ CN_{dw} &= \text{C : N for dead wood (in stem and coarse root)} \end{aligned} \quad (2.693)$$

where all C:N parameters are defined as constants for a given PFT (Table 2.26).

Given values for the parameters in and , total carbon and nitrogen allocation to new growth (CF_{alloc} , gC m⁻² s⁻¹, and NF_{alloc} , gN m⁻² s⁻¹, respectively) can be expressed as functions of new leaf carbon allocation ($CF_{GPP,leaf}$, gC m⁻² s⁻¹):

$$\begin{aligned} CF_{alloc} &= CF_{GPP,leaf} C_{allom} \\ NF_{alloc} &= CF_{GPP,leaf} N_{allom} \end{aligned} \quad (2.694)$$

where

$$C_{allom} = \begin{cases} (1 + g_1)(1 + a_1 + a_3(1 + a_2)) & \text{for woody PFT} \\ 1 + g_1 + a_1(1 + g_1) & \text{for non-woody PFT} \end{cases} \quad (2.695)$$

$$N_{allom} = \begin{cases} \frac{1}{CN_{leaf}} + \frac{a_1}{CN_{fr}} + \frac{a_3 a_4 (1+a_2)}{CN_{lw}} + \\ \frac{a_3 (1-a_4) (1+a_2)}{CN_{dw}} & \text{for woody PFT} \\ \frac{1}{CN_{leaf}} + \frac{a_1}{CN_{fr}} & \text{for non-woody PFT.} \end{cases} \quad (2.696)$$

Since the C:N stoichiometry for new growth allocation is defined, from Eq. , as C_{allom}/N_{allom} , the total carbon available for new growth allocation (CF_{avail_alloc}) can be used to calculate the total plant nitrogen demand for new growth (NF_{plant_demand} , gN m⁻² s⁻¹) as:

$$NF_{plant_demand} = CF_{avail_alloc} \frac{N_{allom}}{C_{allom}}. \quad (2.697)$$

2.19.4 Carbon Allocation to New Growth

There are two carbon pools associated with each plant tissue – one which represents the currently displayed tissue, and another which represents carbon stored for display in a subsequent growth period. The nitrogen pools follow this same organization. The model keeps track of stored carbon according to which tissue type it will eventually be displayed as, and the separation between display in the current timestep and storage for later display depends on the parameter f_{cur} (values 0 to 1). Given $CF_{alloc,leaf}$ and f_{cur} , the allocation fluxes of carbon to display and storage pools (where storage is indicated with $_stor$) for the various tissue types are given as:

$$CF_{alloc,leaf_} = CF_{alloc,leaf_tot} f_{cur} \quad (2.698)$$

$$CF_{alloc,leaf_stor_} = CF_{alloc,leaf_tot} (1 - f_{cur}) \quad (2.699)$$

$$CF_{alloc,froot_} = CF_{alloc,leaf_tot} a_1 f_{cur} \quad (2.700)$$

$$CF_{alloc,froot_stor_} = CF_{alloc,leaf_tot} a_1 (1 - f_{cur}) \quad (2.701)$$

$$CF_{alloc,livestem_} = CF_{alloc,leaf_tot} a_3 a_4 f_{cur} \quad (2.702)$$

$$CF_{alloc,livestem_stor_} = CF_{alloc,leaf_tot} a_3 a_4 (1 - f_{cur}) \quad (2.703)$$

$$CF_{alloc,deadstem_} = CF_{alloc,leaf_tot} a_3 (1 - a_4) f_{cur} \quad (2.704)$$

$$CF_{alloc,deadstem_stor_} = CF_{alloc,leaf_tot} a_3 (1 - a_4) (1 - f_{cur}) \quad (2.705)$$

$$CF_{alloc,livecroo_} = CF_{alloc,leaf_tot} a_2 a_3 a_4 f_{cur} \quad (2.706)$$

$$CF_{alloc,livecroo_stor_} = CF_{alloc,leaf_tot} a_2 a_3 a_4 (1 - f_{cur}) \quad (2.707)$$

$$CF_{alloc,deadcroo_} = CF_{alloc,leaf_tot} a_2 a_3 (1 - a_4) f_{cur} \quad (2.708)$$

$$CF_{alloc,deadcroo_stor_} = CF_{alloc,leaf_tot} a_2 a_3 (1 - a_4) (1 - f_{cur}). \quad (2.709)$$

2.19.5 Nitrogen allocation

The total flux of nitrogen to be allocated is given by the FUN model (Chapter 2.18). This gives a total N to be allocated within a given timestep, N_{supply} . The total N allocated for a given tissue i is the minimum between the supply and the demand:

$$NF_{alloc,i} = \min(NF_{demand,i}, NF_{supply,i}) \quad (2.710)$$

The demand for each tissue, calculated for the tissue to remain on stoichiometry during growth, is:

$$NF_{demand,leaf_} = \frac{CF_{alloc,leaf_tot}}{CN_{leaf}} f_{cur} \quad (2.711)$$

$$NF_{demand,leaf_stor_} = \frac{CF_{alloc,leaf_tot}}{CN_{leaf}} (1 - f_{cur}) \quad (2.712)$$

$$NF_{demand,froot_} = \frac{CF_{alloc,leaf_tot}a_1}{CN_{fr}} f_{cur} \quad (2.713)$$

$$NF_{demand,froot_stor_} = \frac{CF_{alloc,leaf_tot}a_1}{CN_{fr}} (1 - f_{cur}) \quad (2.714)$$

$$NF_{demand,livestem_} = \frac{CF_{alloc,leaf_tot}a_3a_4}{CN_{lw}} f_{cur} \quad (2.715)$$

$$NF_{demand,livestem_stor_} = \frac{CF_{alloc,leaf_tot}a_3a_4}{CN_{lw}} (1 - f_{cur}) \quad (2.716)$$

$$NF_{demand,deadstem_} = \frac{CF_{alloc,leaf_tot}a_3(1-a_4)}{CN_{dw}} f_{cur} \quad (2.717)$$

$$NF_{demand,deadstem_stor_} = \frac{CF_{alloc,leaf_tot}a_3(1-a_4)}{CN_{dw}} (1 - f_{cur}) \quad (2.718)$$

$$NF_{demand,livecroot_} = \frac{CF_{alloc,leaf_tot}a_2a_3a_4}{CN_{lw}} f_{cur} \quad (2.719)$$

$$NF_{demand,livecroot_stor_} = \frac{CF_{alloc,leaf_tot}a_2a_3a_4}{CN_{lw}} (1 - f_{cur}) \quad (2.720)$$

$$NF_{demand,deadcroot_} = \frac{CF_{alloc,leaf_tot}a_2a_3(1-a_4)}{CN_{dw}} f_{cur} \quad (2.721)$$

$$NF_{demand,deadcroot_stor_} = \frac{CF_{alloc,leaf}a_2a_3(1-a_4)}{CN_{dw}} (1 - f_{cur}). \quad (2.722)$$

After each pool's demand is calculated, the total plant N demand is then the sum of each individual pool :math: i corresponding to each tissue:

$$NF_{demand,tot} = \sum_{i=tissues} NF_{demand,i} \quad (2.723)$$

and the total supply for each tissue :math: i is the product of the fractional demand and the total available N, calculated as the term :math: N_{uptake} equal to the sum of the eight N uptake streams described in the FUN model (Chapter 2.18).

$$NF_{alloc,i} = N_{uptake} NF_{demand,i} / NF_{demand,tot} \quad (2.724)$$

2.20 Vegetation Phenology and Turnover

The CLM phenology model consists of several algorithms controlling the transfer of stored carbon and nitrogen out of storage pools for the display of new growth and into litter pools for losses of displayed growth. PFTs are classified into three distinct phenological types that are represented by separate algorithms: an evergreen type, for which some fraction of annual leaf growth persists in the displayed pool for longer than one year; a seasonal deciduous type with a single growing season per year, controlled mainly by temperature and daylength; and a stress-deciduous type with the potential for multiple growing seasons per year, controlled by temperature and soil moisture conditions.

The three phenology types share a common set of control variables. The calculation of the phenology fluxes is generalized, operating identically for all three phenology types, given a specification of the common control variables. The following sections describe first the general flux parameterization, followed by the algorithms for setting the control parameters for the three phenology types.

2.20.1 General Phenology Flux Parameterization

Fluxes of carbon and nitrogen from storage pools and into displayed tissue pools pass through a special transfer pool (denoted *_xfer*), maintained as a separate state variable for each tissue type. Storage (*_stor*) and transfer (*_xfer*) pools are maintained separately to reduce the complexity of accounting for transfers into and out of storage over the course of a single growing season.

14.1.1 Onset Periods

The deciduous phenology algorithms specify the occurrence of onset growth periods (Figure 14.1). Carbon fluxes from the transfer pools into displayed growth are calculated during these periods as:

$$CF_{leaf_xfer,leaf} = r_{xfer_on} CS_{leaf_xfer} \quad (2.725)$$

$$CF_{froot_xfer,froot} = r_{xfer_on} CS_{froot_xfer} \quad (2.726)$$

$$CF_{livesetem_xfer,livesetem} = r_{xfer_on} CS_{livesetem_xfer} \quad (2.727)$$

$$CF_{deadstem_xfer,deadstem} = r_{xfer_on} CS_{deadstem_xfer} \quad (2.728)$$

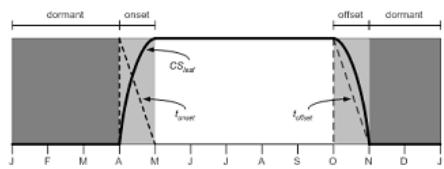


Figure 2.15: Example of annual phenology cycle for seasonal deciduous.

$$CF_{livecroo_xfer,livecroo} = r_{xfer_on} CS_{livecroo_xfer} \quad (2.729)$$

$$CF_{deadcroo_xfer,deadcroo} = r_{xfer_on} CS_{deadcroo_xfer}, \quad (2.730)$$

with corresponding nitrogen fluxes:

$$NF_{leaf_xfer,leaf} = r_{xfer_on} NS_{leaf_xfer} \quad (2.731)$$

$$NF_{froot_xfer,froot} = r_{xfer_on} NS_{froot_xfer} \quad (2.732)$$

$$NF_{livestem_xfer,livestem} = r_{xfer_on} NS_{livestem_xfer} \quad (2.733)$$

$$NF_{deadstem_xfer,deadstem} = r_{xfer_on} NS_{deadstem_xfer} \quad (2.734)$$

$$NF_{livecroo_xfer,livecroo} = r_{xfer_on} NS_{livecroo_xfer} \quad (2.735)$$

$$NF_{deadcroo_xfer,deadcroo} = r_{xfer_on} NS_{deadcroo_xfer}, \quad (2.736)$$

where CF is the carbon flux, CS is stored carbon, NF is the nitrogen flux, NS is stored nitrogen, r_{xfer_on} (s^{-1}) is a time-varying rate coefficient controlling flux out of the transfer pool:

$$r_{xfer_on} = \begin{cases} 2/t_{onset} & \text{for } t_{onset} \neq \Delta t \\ 1/\Delta t & \text{for } t_{onset} = \Delta t \end{cases} \quad (2.737)$$

and t_{onset} (s) is the number of seconds remaining in the current phenology onset growth period (Figure 14.1). The form of Eq. (2.737) produces a flux from the transfer pool which declines linearly over the onset growth period, approaching zero flux in the final timestep.

14.1.2 Offset Periods

The deciduous phenology algorithms also specify the occurrence of litterfall during offset periods. In contrast to the onset periods, only leaf and fine root state variables are subject to litterfall fluxes. Carbon fluxes from display pools into litter are calculated during these periods as:

$$CF_{leaf,litter}^n = \begin{cases} CF_{leaf,litter}^{n-1} + r_{xfer_off} (CS_{leaf} - CF_{leaf,litter}^{n-1} t_{offset}) & \text{for } t_{offset} \neq \Delta t \\ (CS_{leaf}/\Delta t) + CF_{alloc,leaf} & \text{for } t_{offset} = \Delta t \end{cases} \quad (2.738)$$

$$CF_{froot,litter}^n = \begin{cases} CF_{froot,litter}^{n-1} + r_{xfer_off} (CS_{froot} - CF_{froot,litter}^{n-1} t_{offset}) & \text{for } t_{offset} \neq \Delta t \\ (CS_{froot}/\Delta t) + CF_{alloc,froot} & \text{for } t_{offset} = \Delta t \end{cases} \quad (2.739)$$

$$r_{xfer_off} = \frac{2\Delta t}{t_{offset}^2} \quad (2.740)$$

where superscripts n and $n-1$ refer to fluxes on the current and previous timesteps, respectively. The rate coefficient r_{xfer_off} varies with time to produce a linearly increasing litterfall rate throughout the offset period, and the special case for fluxes in the final litterfall timestep ($t_{offset} = \Delta t$) ensures that all of the displayed growth is sent to the litter pools for deciduous plant types.

Corresponding nitrogen fluxes during litterfall take into account retranslocation of nitrogen out of the displayed leaf pool prior to litterfall ($NF_{leaf,retrans}$, gN m⁻² s⁻¹). Retranslocation of nitrogen out of fine roots is assumed to be negligible. The fluxes are:

$$NF_{leaf,litter} = CF_{leaf,litter}/CN_{leaf_litter} \quad (2.741)$$

$$NF_{froot,litter} = CF_{leaf,litter}/CN_{froot} \quad (2.742)$$

$$NF_{leaf,retrans} = (CF_{leaf,litter}/CN_{leaf}) - NF_{leaf,litter}. \quad (2.743)$$

where CN is C:N.

14.1.3 Background Onset Growth

The stress-deciduous phenology algorithm includes a provision for the case when stress signals are absent, and the vegetation shifts from a deciduous habit to an evergreen habit, until the next occurrence of an offset stress trigger. In that case, the regular onset flux mechanism is switched off and a background onset growth algorithm is invoked ($r_{bgtr} > 0$). During this period, small fluxes of carbon and nitrogen from the storage pools into the associated transfer pools are calculated on each time step, and the entire contents of the transfer pool are added to the associated displayed growth pool on each time step. The carbon fluxes from transfer to display pools under these conditions are:

$$CF_{leaf_xfer,leaf} = CS_{leaf_xfer}/\Delta t \quad (2.744)$$

$$CF_{froot_xfer,froot} = CS_{froot_xfer}/\Delta t \quad (2.745)$$

$$CF_{livestem_xfer,livestem} = CS_{livestem_xfer}/\Delta t \quad (2.746)$$

$$CF_{deadstem_xfer,deadstem} = CS_{deadstem_xfer}/\Delta t \quad (2.747)$$

$$CF_{livecroo_xfer,livecroo} = CS_{livecroo_xfer}/\Delta t \quad (2.748)$$

$$CF_{deadcroo_xfer,deadcroo} = CS_{deadcroo_xfer}/\Delta t, \quad (2.749)$$

and the corresponding nitrogen fluxes are:

$$NF_{leaf_xfer,leaf} = NS_{leaf_xfer}/\Delta t \quad (2.750)$$

$$NF_{froot_xfer,froot} = NS_{froot_xfer}/\Delta t \quad (2.751)$$

$$NF_{livestem_xfer,livestem} = NS_{livestem_xfer}/\Delta t \quad (2.752)$$

$$NF_{deadstem_xfer,deadstem} = NS_{deadstem_xfer}/\Delta t \quad (2.753)$$

$$NF_{livecroot_xfer,livecroot} = NS_{livecroot_xfer}/\Delta t \quad (2.754)$$

$$NF_{deadcroot_xfer,deadcroot} = NS_{deadcroot_xfer}/\Delta t. \quad (2.755)$$

14.1.4 Background Litterfall

Both evergreen and stress-deciduous phenology algorithms can specify a litterfall flux that is not associated with a specific offset period, but which occurs instead at a slow rate over an extended period of time, referred to as background litterfall. For evergreen types the background litterfall is the only litterfall flux. For stress-deciduous types either the offset period litterfall or the background litterfall mechanism may be active, but not both at once. Given a specification of the background litterfall rate (r_{bgf} , s^{-1}), litterfall carbon fluxes are calculated as

$$CF_{leaf,litter} = r_{bgf} CS_{leaf} \quad (2.756)$$

$$CS_{froot,litter} = r_{bgf} CS_{froot}, \quad (2.757)$$

with corresponding nitrogen litterfall and retranslocation fluxes:

$$NF_{leaf,litter} = CF_{leaf,litter}/CN_{leaf_litter} \quad (2.758)$$

$$NF_{froot,litter} = CF_{froot,litter}/CN_{froot} \quad (2.759)$$

$$NF_{leaf,retrans} = (CF_{leaf,litter}/CN_{leaf}) - NF_{leaf,litter}. \quad (2.760)$$

14.1.5 Livewood Turnover

The conceptualization of live wood vs. dead wood fractions for stem and coarse root pools is intended to capture the difference in maintenance respiration rates between these two physiologically distinct tissue types. Unlike displayed pools for leaf and fine root, which are lost to litterfall, live wood cells reaching the end of their lifespan are retained as a part of the dead woody structure of stems and coarse roots. A mechanism is therefore included in the phenology routine to effect the transfer of live wood to dead wood pools, which also takes into account the different nitrogen concentrations typical of these tissue types.

A live wood turnover rate (r_{lw} , s⁻¹) is defined as

$$r_{lw} = p_{lw} / (365 \cdot 86400) \quad (2.761)$$

where $p_{lw} = 0.7$ is the assumed annual live wood turnover fraction. Carbon fluxes from live to dead wood pools are:

$$CF_{livesetem,deadstem} = CS_{livesetem} r_{lw} \quad (2.762)$$

$$CF_{livecroo,deadcroo} = CS_{livecroo} r_{lw}, \quad (2.763)$$

and the associated nitrogen fluxes, including retranslocation of nitrogen out of live wood during turnover, are:

$$NF_{livesetem,deadstem} = CF_{livesetem,deadstem} / CN_{dw} \quad (2.764)$$

$$NF_{livesetem,retrans} = (CF_{livesetem,deadstem} / CN_{lw}) - NF_{livesetem,deadstem} \quad (2.765)$$

$$NF_{livecroo,deadcroo} = CF_{livecroo,deadcroo} / CN_{dw} \quad (2.766)$$

$$NF_{livecroo,retrans} = (CF_{livecroo,deadcroo} / CN_{lw}) - NF_{livecroo,deadcroo}. \quad (2.767)$$

2.20.2 Evergreen Phenology

The evergreen phenology algorithm is by far the simplest of the three possible types. It is assumed for all evergreen types that all carbon and nitrogen allocated for new growth in the current timestep goes immediately to the displayed growth pools (i.e. $ff_{cur} = 1.0$ (Chapter 13)). As such, there is never an accumulation of carbon or nitrogen in the storage or transfer pools, and so the onset growth and background onset growth mechanisms are never invoked for this type. Litterfall is specified to occur only through the background litterfall mechanism – there are no distinct periods of litterfall for evergreen types, but rather a continuous (slow) shedding of foliage and fine roots. This is an obvious area for potential improvements in the model, since it is known, at least for evergreen needleleaf trees in the temperate and boreal zones, that there are distinct periods of higher and lower leaf litterfall (Ferrari, 1999; Gholz et al., 1985). The rate of background litterfall (r_{bgf} , section 14.1.4) depends on the specified leaf longevity (τ_{leaf} , y), as

$$r_{bgf} = \frac{1}{\tau_{leaf} \cdot 365 \cdot 86400}. \quad (2.768)$$

2.20.3 Seasonal-Deciduous Phenology

The seasonal-deciduous phenology algorithm derives directly from the treatment used in the offline model Biome-BGC v. 4.1.2, (Thornton et al., 2002), which in turn is based on the parameterizations for leaf onset and offset for temperate deciduous broadleaf forest from White et al. (1997). Initiation of leaf onset is triggered when a common degree-day summation exceeds a critical value, and leaf litterfall is initiated when daylength is shorter than a critical value. Because of the dependence on daylength, the seasonal deciduous phenology algorithm is only valid for latitudes outside of the tropical zone, defined here as $|latitude| > 19.5^\circ$. Neither the background onset nor background litterfall mechanism is invoked for the seasonal-deciduous phenology algorithm. The algorithm allows a maximum of one onset period and one offset period each year.

The algorithms for initiation of onset and offset periods use the winter and summer solstices as coordination signals. The period between winter and summer solstice is identified as $dayl_n > dayl_{n-1}$, and the period between summer and winter solstice is identified as $dayl_n < dayl_{n-1}$, where $dayl_n$ and $dayl_{n-1}$ are the day length(s) calculated for the current and previous timesteps, respectively, using

$$dayl = 2 \cdot 13750.9871 \cdot \arccos\left(\frac{-\sin(lat)\sin(decl)}{\cos(lat)\cos(decl)}\right), \quad (2.769)$$

where lat and $decl$ are the latitude and solar declination (radians), respectively, and the factor 13750.9871 is the number of seconds per radian of hour-angle.

14.3.1 Seasonal-Deciduous Onset Trigger

The onset trigger for the seasonal-deciduous phenology algorithm is based on an accumulated growing-degree-day approach (White et al., 1997). The growing-degree-day summation (GDD_{sum}) is initiated ($GDD_{sum} = 0$) when the phenological state is dormant and the model timestep crosses the winter solstice. Once these conditions are met, GDD_{sum} is updated on each timestep as

$$GDD_{sum}^n = \begin{cases} GDD_{sum}^{n-1} + (T_{s,3} - TKFRZ) f_{day} & \text{for } T_{s,3} > TKFRZ \\ GDD_{sum}^{n-1} & \text{for } T_{s,3} \leq TKFRZ \end{cases} \quad (2.770)$$

where $T_{s,3}$ (K) is the temperature of the third soil layer, and $f_{day} = \Delta t / 86400$. The onset period is initiated if $GDD_{sum} > GDD_{sum_crit}$, where

$$GDD_{sum_crit} = \exp(4.8 + 0.13(T_{2m,ann_avg} - TKFRZ)) \quad (2.771)$$

and where T_{2m,ann_avg} (K) is the annual average of the 2m air temperature, and TKFRZ is the freezing point of water (273.15 K). The following control variables are set when a new onset growth period is initiated:

$$GDD_{sum} = 0 \quad (2.772)$$

$$t_{onset} = 86400 \cdot n_{days_on}, \quad (2.773)$$

where n_{days_on} is set to a constant value of 30 days. Fluxes from storage into transfer pools occur in the timestep when a new onset growth period is initiated. Carbon fluxes are:

$$CF_{leaf_stor,leaf_xfer} = f_{stor,xfer} CS_{leaf_stor} / \Delta t \quad (2.774)$$

$$CF_{froot_stor,froot_xfer} = f_{stor,xfer} CS_{froot_stor} / \Delta t \quad (2.775)$$

$$CF_{livestem_stor,livestem_xfer} = f_{stor,xfer} CS_{livestem_stor} / \Delta t \quad (2.776)$$

$$CF_{deadstem_stor,deadstem_xfer} = f_{stor,xfer} CS_{deadstem_stor} / \Delta t \quad (2.777)$$

$$CF_{livecroo_stor,livecroo_xfer} = f_{stor,xfer} CS_{livecroo_stor} / \Delta t \quad (2.778)$$

$$CF_{deadcroo_stor,deadcroo_xfer} = f_{stor,xfer} CS_{deadcroo_stor} / \Delta t \quad (2.779)$$

$$CF_{gresp_stor,gresp_xfer} = f_{stor,xfer} CS_{gresp_stor} / \Delta t \quad (2.780)$$

and the associated nitrogen fluxes are:

$$NF_{leaf_stor,leaf_xfer} = f_{stor,xfer} NS_{leaf_stor} / \Delta t \quad (2.781)$$

$$NF_{froot_stor,froot_xfer} = f_{stor,xfer} NS_{froot_stor} / \Delta t \quad (2.782)$$

$$NF_{livestem_stor,livestem_xfer} = f_{stor,xfer} NS_{livestem_stor} / \Delta t \quad (2.783)$$

$$NF_{deadstem_stor,deadstem_xfer} = f_{stor,xfer} NS_{deadstem_stor} / \Delta t \quad (2.784)$$

$$NF_{livecroo_stor,livecroo_xfer} = f_{stor,xfer} NS_{livecroo_stor} / \Delta t \quad (2.785)$$

$$NF_{deadcroo_stor,deadcroo_xfer} = f_{stor,xfer} NS_{deadcroo_stor} / \Delta t \quad (2.786)$$

where $f_{stor,xfer}$ is the fraction of current storage pool moved into the transfer pool for display over the incipient onset period. This fraction is set to 0.5, based on the observation that seasonal deciduous trees are capable of replacing their canopies from storage reserves in the event of a severe early-season disturbance such as frost damage or defoliation due to insect herbivory.

If the onset criterion ($GDD_{sum} > GDD_{sum_crit}$) is not met before the summer solstice, then GDD_{sum} is set to 0.0 and the growing-degree-day accumulation will not start again until the following winter solstice. This mechanism prevents the initiation of very short growing seasons late in the summer in cold climates. The onset counter is decremented on each time step after initiation of the onset period, until it reaches zero, signaling the end of the onset period:

$$t_{onset}^n = t_{onset}^{n-1} - \Delta t \quad (2.787)$$

14.3.2 Seasonal-Deciduous Offset Trigger

After the completion of an onset period, and once past the summer solstice, the offset (litterfall) period is triggered when daylength is shorter than 39300 s. The offset counter is set at the initiation of the offset period: $t_{offset} = 86400 \cdot n_{days_off}$, where n_{days_off} is set to a constant value of 15 days. The offset counter is decremented on each time step after initiation of the offset period, until it reaches zero, signaling the end of the offset period:

$$t_{offset}^n = t_{offset}^{n-1} - \Delta t \quad (2.788)$$

2.20.4 Stress-Deciduous Phenology

The stress-deciduous phenology algorithm was developed specifically for the CLM based in part on the grass phenology model proposed by White et al. (1997). The algorithm handles phenology for vegetation types such as grasses and tropical drought-deciduous trees that respond to both cold and drought-stress signals, and that can have multiple growing seasons per year. The algorithm also allows for the possibility that leaves might persist year-round in the absence of a suitable stress trigger. In that case the phenology switches to an evergreen habit, maintaining a marginally-deciduous leaf longevity (one year) until the occurrence of the next stress trigger.

14.4.1 Stress-Deciduous Onset Triggers

In climates that are warm year-round, onset triggering depends on soil water availability. At the beginning of a dormant period (end of previous offset period), an accumulated soil water index (SWI_{sum} , d) is initialized ($SWI_{sum} = 0$), with subsequent accumulation calculated as:

$$SWI_{sum}^n = \begin{cases} SWI_{sum}^{n-1} + f_{day} & \text{for } \Psi_{s,3} \geq \Psi_{onset} \\ SWI_{sum}^{n-1} & \text{for } \Psi_{s,3} < \Psi_{onset} \end{cases} \quad (2.789)$$

where $\Psi_{s,3}$ is the soil water potential (MPa) in the third soil layer and $\Psi_{onset} = -0.6\text{ MPa}$ is the onset soil water potential threshold. Onset triggering is possible once $SWI_{sum} > 15$. To avoid spurious onset triggering due to soil moisture in the third soil layer exceeding the threshold due only to soil water suction of water from deeper in the soil column, an additional precipitation trigger is included which requires at least 20 mm of rain over the previous 10 days (Dahlin et al., 2015). If the cold climate growing degree-day accumulator is not active at the time when the soil moisture and precipitation thresholds are reached (see below), and if the daylength is greater than 6 hours, then onset is triggered. Except as noted below, SWI_{sum} continues to accumulate according to Eq. (2.789) during the dormant period if the daylength criterion prevents onset triggering, and onset is then triggered at the timestep when daylength exceeds 6 hours.

In climates with a cold season, onset triggering depends on both accumulated soil temperature summation and adequate soil moisture. At the beginning of a dormant period a freezing day accumulator (FD_{sum} , d) is initialized ($FD_{sum} = 0$), with subsequent accumulation calculated as:

$$FD_{sum}^n = \begin{cases} FD_{sum}^{n-1} + f_{day} & \text{for } T_{s,3} > TKFRZ \\ FD_{sum}^{n-1} & \text{for } T_{s,3} \leq TKFRZ \end{cases} \quad (2.790)$$

If $FD_{sum} > 15$ during the dormant period, then a cold-climate onset triggering criterion is introduced, following exactly the growing degree-day summation (GDD_{sum}) logic of Eqs. (2.770) and (2.771). At that time SWI_{sum} is reset ($SWI_{sum} = 0$). Onset triggering under these conditions depends on meeting all three of the following criteria: $SWI_{sum} > 15$, $GDD_{sum} > GDD_{sum_crit}$, and daylength greater than 6 hrs.

The following control variables are set when a new onset growth period is initiated: $SWI_{sum} = 0$, $FD_{sum} = 0$, $GDD_{sum} = 0$, $n_{days_active} = 0$, and $t_{onset} = 86400 \cdot n_{days_on}$, where n_{days_on} is set to a constant value of 30 days. Fluxes from storage into transfer pools occur in the timestep when a new onset growth period is initiated, and are handled identically to Eqs. (2.774) -(2.780) for carbon fluxes, and to Eqs. (2.781) - (2.786) for nitrogen fluxes.

The onset counter is decremented on each time step after initiation of the onset period, until it reaches zero, signaling the end of the onset period:

$$t_{onset}^n = t_{onset}^{n-1} - \Delta t \quad (2.791)$$

14.4.2 Stress-Deciduous Offset Triggers

Any one of the following three conditions is sufficient to initiate an offset period for the stress-deciduous phenology algorithm: sustained period of dry soil, sustained period of cold temperature, or daylength shorter than 6 hours. Offset triggering due to dry soil or cold temperature conditions is only allowed once the most recent onset period is complete. Dry soil condition is evaluated with an offset soil water index accumulator ($OSWI_{sum}$, d). To test for a sustained period of dry soils, this control variable can increase or decrease, as follows:

$$OSWI_{sum}^n = \begin{cases} OSWI_{sum}^{n-1} + f_{day} & \text{for } \Psi_{s,3} \leq \Psi_{offset} \\ \max(OSWI_{sum}^{n-1} - f_{day}, 0) & \text{for } \Psi_{s,3} > \Psi_{onset} \end{cases} \quad (2.792)$$

where $\Psi_{offset} = -2 MPa$ is the offset soil water potential threshold. An offset period is triggered if the previous onset period is complete and $OSWI_{sum} \geq OSWI_{sum_crit}$, where $OSWI_{sum_crit} = 15$.

The cold temperature trigger is calculated with an offset freezing day accumulator (OFD_{sum} , d). To test for a sustained period of cold temperature, this variable can increase or decrease, as follows:

$$OFD_{sum}^n = \begin{cases} OFD_{sum}^{n-1} + f_{day} & \text{for } T_{s,3} \leq TKFRZ \\ \max(OFD_{sum}^{n-1} - f_{day}, 0) & \text{for } T_{s,3} > TKFRZ \end{cases} \quad (2.793)$$

An offset period is triggered if the previous onset period is complete and $OFD_{sum} > OFD_{sum_crit}$, where $OFD_{sum_crit} = 15$.

The offset counter is set at the initiation of the offset period: $t_{offset} = 86400 \cdot n_{days_off}$, where n_{days_off} is set to a constant value of 15 days. The offset counter is decremented on each time step after initiation of the offset period, until it reaches zero, signaling the end of the offset period:

$$t_{offset}^n = t_{offset}^{n-1} - \Delta t \quad (2.794)$$

14.4.3 Stress-Deciduous: Long Growing Season

Under conditions when the stress-deciduous conditions triggering offset are not met for one year or longer, the stress-deciduous algorithm shifts toward the evergreen behavior. This can happen in cases where a stress-deciduous vegetation type is assigned in a climate where suitably strong stresses occur less frequently than once per year. This condition is evaluated by tracking the number of days since the beginning of the most recent onset period (n_{days_active} , d). At the end of an offset period n_{days_active} is reset to 0. A long growing season control variable (LGS , range 0 to 1) is calculated as:

$$LGS = \begin{cases} 0 & \text{for } n_{days_active} < 365 \\ (n_{days_active}/365) - 1 & \text{for } 365 \leq n_{days_active} < 730 \\ 1 & \text{for } n_{days_active} \geq 730 \end{cases} \quad (2.795)$$

The rate coefficient for background litterfall (r_{bgf} , s⁻¹) is calculated as a function of LGS :

$$r_{bgf} = \frac{LGS}{\tau_{leaf} \cdot 365 \cdot 86400} \quad (2.796)$$

where τ_{leaf} is the leaf longevity. The result is a shift to continuous litterfall as n_{days_active} increases from 365 to 730. When a new offset period is triggered r_{bgf} is set to 0.

The rate coefficient for background onset growth from the transfer pools (r_{bgtr} , s^{-1}) also depends on LGS , as:

$$r_{bgtr} = \frac{LGS}{365 \cdot 86400}. \quad (2.797)$$

On each timestep with $r_{bgtr} \neq 0$, carbon fluxes from storage to transfer pools are calculated as:

$$CF_{leaf_stor,leaf_xfer} = CS_{leaf_stor} r_{bgtr} \quad (2.798)$$

$$CF_{froot_stor,froot_xfer} = CS_{froot_stor} r_{bgtr} \quad (2.799)$$

$$CF_{livesetm_stor,livesetm_xfer} = CS_{livesetm_stor} r_{bgtr} \quad (2.800)$$

$$CF_{deadstem_stor,deadstem_xfer} = CS_{deadstem_stor} r_{bgtr} \quad (2.801)$$

$$CF_{livecroot_stor,livecroot_xfer} = CS_{livecroot_stor} r_{bgtr} \quad (2.802)$$

$$CF_{deadcroot_stor,deadcroot_xfer} = CS_{deadcroot_stor} r_{bgtr}, \quad (2.803)$$

with corresponding nitrogen fluxes:

$$NF_{leaf_stor,leaf_xfer} = NS_{leaf_stor} r_{bgtr} \quad (2.804)$$

$$NF_{froot_stor,froot_xfer} = NS_{froot_stor} r_{bgtr} \quad (2.805)$$

$$NF_{livesetm_stor,livesetm_xfer} = NS_{livesetm_stor} r_{bgtr} \quad (2.806)$$

$$NF_{deadstem_stor,deadstem_xfer} = NS_{deadstem_stor} r_{bgtr} \quad (2.807)$$

$$NF_{livecroot_stor,livecroot_xfer} = NS_{livecroot_stor} r_{bgtr} \quad (2.808)$$

$$NF_{deadcroot_stor,deadcroot_xfer} = NS_{deadcroot_stor} r_{bgtr}. \quad (2.809)$$

The result, in conjunction with the treatment of background onset growth, is a shift to continuous transfer from storage to display pools at a rate that would result in complete turnover of the storage pools in one year at steady state, once LGS reaches 1 (i.e. after two years without stress-deciduous offset conditions). If and when conditions cause stress-deciduous triggering again, r_{bgtr} is rest to 0.

2.20.5 Litterfall Fluxes Merged to the Column Level

CLM uses three litter pools, defined on the basis of commonly measured chemical fractionation of fresh litter into labile (LIT1 = hot water and alcohol soluble fraction), cellulose/hemicellulose (LIT2 = acid soluble fraction) and remaining material, referred to here for convenience as lignin (LIT3 = acid insoluble fraction) (Aber et al., 1990; Taylor et al., 1989). While multiple plant functional types can coexist on a single CLM soil column, each soil column includes a single instance of the litter pools. Fluxes entering the litter pools due to litterfall are calculated using a weighted average of the fluxes originating at the PFT level. Carbon fluxes are calculated as:

$$CF_{leaf,lit1} = \sum_{p=0}^{npfts} CF_{leaf,litter} f_{lab_leaf,p} wcol_p \quad (2.810)$$

$$CF_{leaf,lit2} = \sum_{p=0}^{npfts} CF_{leaf,litter} f_{cel_leaf,p} wcol_p \quad (2.811)$$

$$CF_{leaf,lit3} = \sum_{p=0}^{npfts} CF_{leaf,litter} f_{lig_leaf,p} wcol_p \quad (2.812)$$

$$CF_{froot,lit1} = \sum_{p=0}^{npfts} CF_{froot,litter} f_{lab_froot,p} wcol_p \quad (2.813)$$

$$CF_{froot,lit2} = \sum_{p=0}^{npfts} CF_{froot,litter} f_{cel_froot,p} wcol_p \quad (2.814)$$

$$CF_{froot,lit3} = \sum_{p=0}^{npfts} CF_{froot,litter} f_{lig_froot,p} wcol_p, \quad (2.815)$$

where $f_{lab_leaf,p}$, $f_{cel_leaf,p}$, and $f_{lig_leaf,p}$ are the labile, cellulose/hemicellulose, and lignin fractions of leaf litter for PFT p , $f_{lab_froot,p}$, $f_{cel_froot,p}$, and $f_{lig_froot,p}$ are the labile, cellulose/hemicellulose, and lignin fractions of fine root litter for PFT p , $wcol_p$ is the weight relative to the column for PFT p , and p is an index through the plant functional types occurring on a column. Nitrogen fluxes to the litter pools are assumed to follow the C:N of the senescent tissue, and so are distributed using the same fractions used for carbon fluxes:

$$NF_{leaf,lit1} = \sum_{p=0}^{npfts} NF_{leaf,litter} f_{lab_leaf,p} wcol_p \quad (2.816)$$

$$NF_{leaf,lit2} = \sum_{p=0}^{npfts} NF_{leaf,litter} f_{cel_leaf,p} wcol_p \quad (2.817)$$

$$NF_{leaf,lit3} = \sum_{p=0}^{npfts} NF_{leaf,litter} f_{lig_leaf,p} wcol_p \quad (2.818)$$

$$NF_{froot,lit1} = \sum_{p=0}^{npfts} NF_{froot,litter} f_{lab_froot,p} wcol_p \quad (2.819)$$

$$NF_{froot,lit2} = \sum_{p=0}^{npfts} NF_{froot,litter} f_{cel_froot,p} wcol_p \quad (2.820)$$

$$NF_{froot,lit3} = \sum_{p=0}^{npfts} NF_{froot,litter} f_{lig_froot,p} wcol_p. \quad (2.821)$$

2.21 Decomposition

Decomposition of fresh litter material into progressively more recalcitrant forms of soil organic matter is represented in CLM as a cascade of k_{tras} transformations between m_{pool} decomposing coarse woody debris (CWD), litter, and soil organic matter (SOM) pools, each defined at n_{lev} vertical levels. CLM allows the user to define, at compile time, between 2 contrasting hypotheses of decomposition as embodied by two separate decomposition submodels: the CLM-CN pool structure used in CLM4.0, or a second pool structure, characterized by slower decomposition rates, based on the fCentury model (Parton et al. 1988). In addition, the user can choose, at compile time, whether to allow n_{lev} to equal 1, as in CLM4.0, or to equal the number of soil levels used for the soil hydrological and thermal calculations (see Section 2.2.2 for soil layering).

Model is structured to allow different representations of the soil C and N decomposition cascade, as well as a vertically-explicit treatment of soil biogeochemistry.

For the single-level model structure, the fundamental equation for carbon balance of the decomposing pools is:

$$\frac{\partial C_i}{\partial t} = R_i + \sum_{j \neq i} (i - r_j) T_{ji} k_j C_j - k_i C_i \quad (2.822)$$

where C_i is the carbon content of pool i , R_i are the carbon inputs from plant tissues directly to pool i (only non-zero for CWD and litter pools), k_i is the decay constant of pool i ; T_{ji} is the fraction of carbon directed from pool j to pool i with fraction r_j lost as a respiration flux along the way.

Adding the vertical dimension to the decomposing pools changes the balance equation to the following:

$$\begin{aligned} \frac{\partial C_i(z)}{\partial t} &= R_i(z) + \sum_{j \neq i} (1 - r_j) T_{ji} k_j(z) C_j(z) - k_i(z) C_i(z) \\ &+ \frac{\partial}{\partial z} (D(z) \frac{\partial C_i}{\partial z}) + \frac{\partial}{\partial z} (A(z) C_i) \end{aligned} \quad (2.823)$$

where $C_i(z)$ is now defined at each model level, and in volumetric (gC m^{-3}) rather than areal (gC m^{-2}) units, along with $R_i(z)$ and $k_j(z)$. In addition, vertical transport is handled by the last two terms, for diffusive and advective transport. In the base model, advective transport is set to zero, leaving only a diffusive flux with diffusivity $D(z)$ defined for all decomposing carbon and nitrogen pools. Further discussion of the vertical distribution of carbon inputs $R_i(z)$, vertical turnover times $k_j(z)$, and vertical transport $D(z)$ is below. Discussion of the vertical model and analysis of both decomposition structures is in [Koven et al. \(2013\)](#).

2.21.1 CLM-CN Pool Structure, Rate Constants and Parameters

The CLM-CN structure in CLM45 uses three state variables for fresh litter and four state variables for soil organic matter (SOM). The masses of carbon and nitrogen in the live microbial community are not modeled explicitly, but the activity of these organisms is represented by decomposition fluxes transferring mass between the litter and SOM pools, and heterotrophic respiration losses associated with these transformations. The litter and SOM pools in CLM-CN are arranged as a converging cascade (Figure 15.2), derived directly from the implementation in Biome-BGC v4.1.2 (Thornton et al. 2002; Thornton and Rosenbloom, 2005).

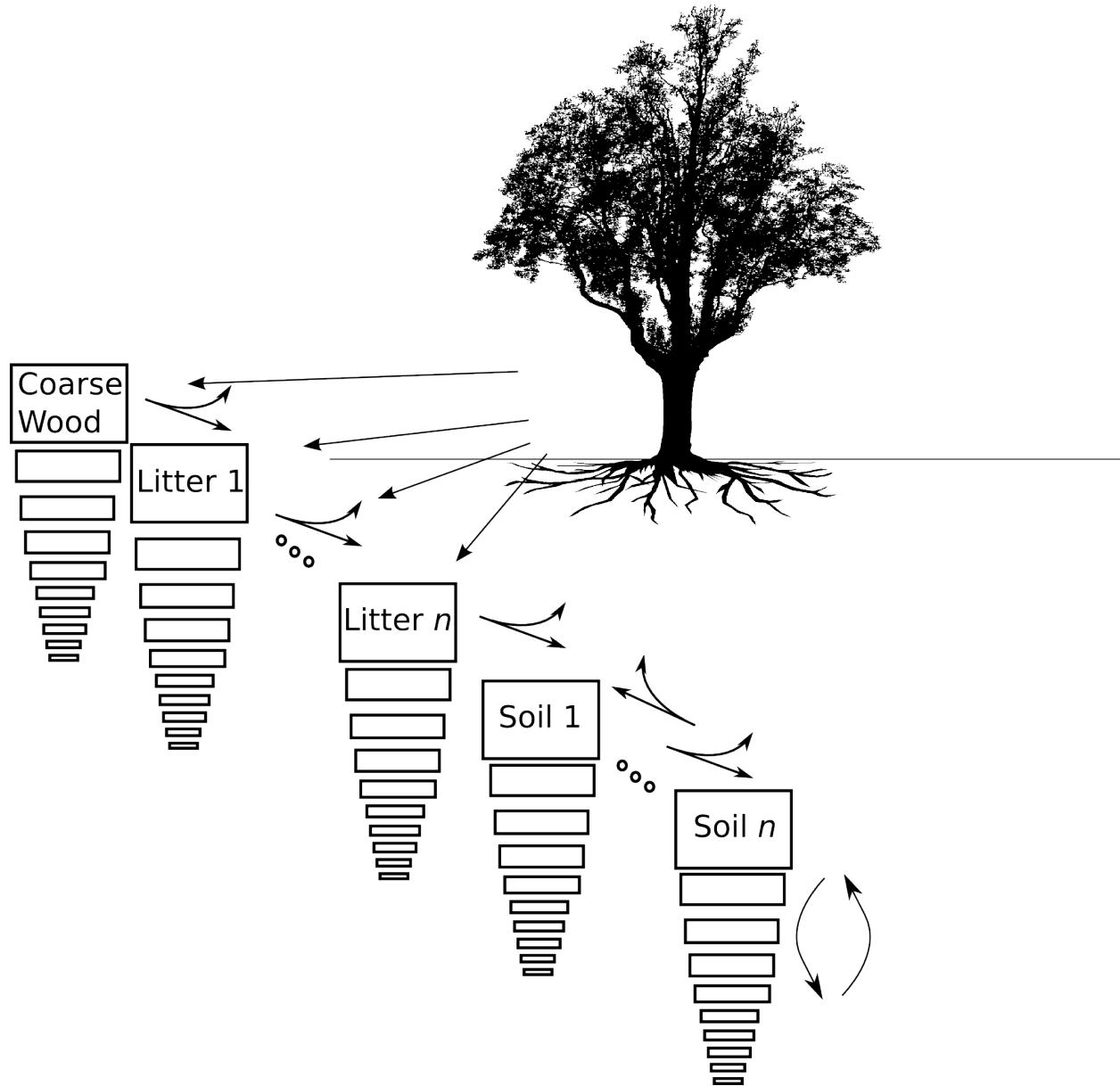


Figure 2.16: Schematic of decomposition model in CLM.

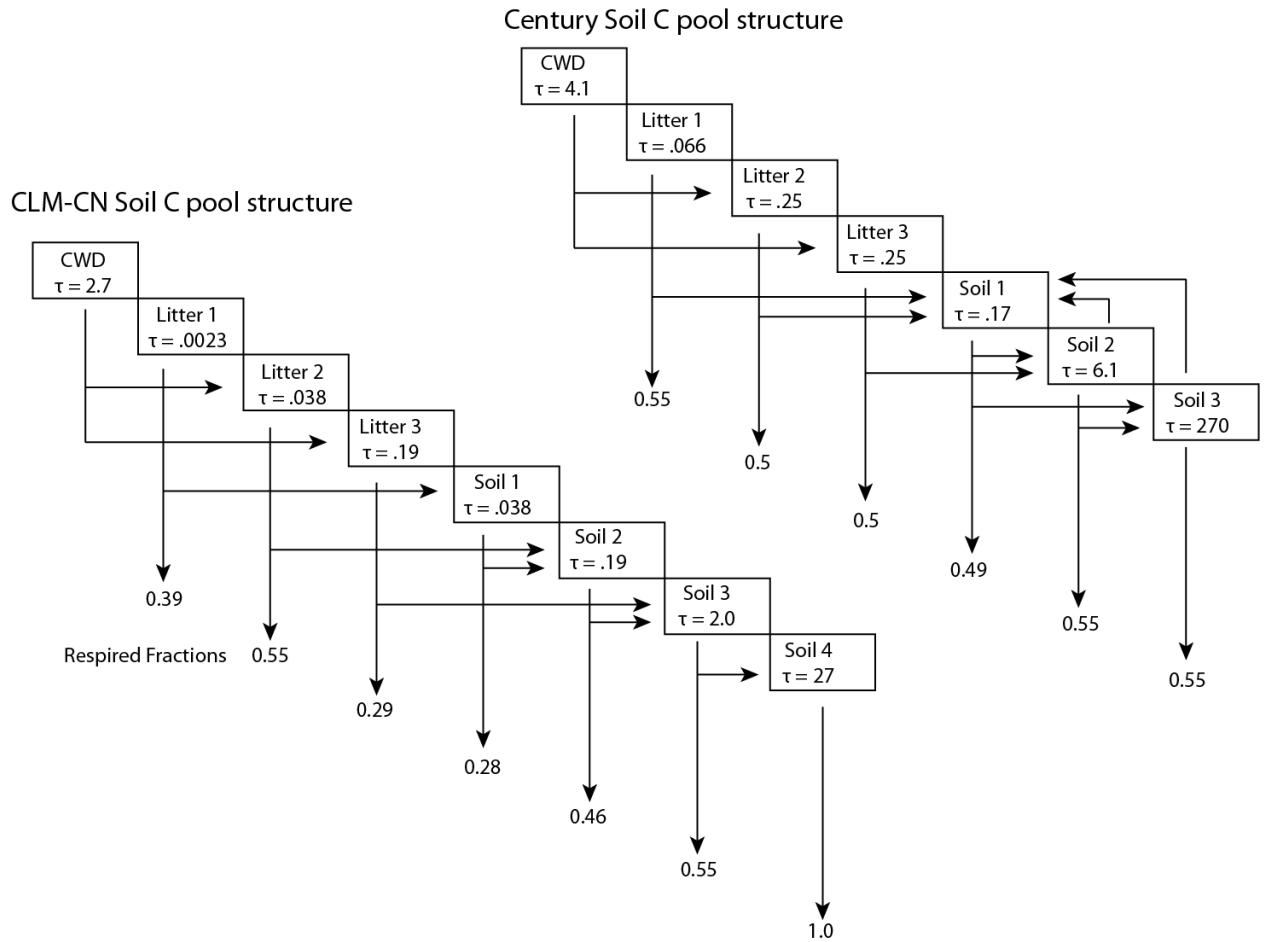


Figure 2.17: Pool structure, transitions, respiration fractions (numbers at end of arrows), and turnover times (numbers in boxes) for the 2 alternate soil decomposition models included in CLM.

Model parameters are estimated based on a synthesis of microcosm decomposition studies using radio-labeled substrates (Degens and Sparling, 1996; Ladd et al. 1992; Martin et al. 1980; Mary et al. 1993; Saggar et al. 1994; Sørensen, 1981; van Veen et al. 1984). Multiple exponential models are fitted to data from the microcosm studies to estimate exponential decay rates and respiration fractions (Thornton, 1998). The microcosm experiments used for parameterization were all conducted at constant temperature and under moist conditions with relatively high mineral nitrogen concentrations, and so the resulting rate constants are assumed not limited by the availability of water or mineral nitrogen. Table 2.27 lists the base decomposition rates for each litter and SOM pool, as well as a base rate for physical fragmentation for the coarse woody debris pool (CWD).

Table 2.27: Decomposition rate constants for litter and SOM pools, C:N ratios, and acceleration parameters (see section 15.8 for explanation) for the CLM-CN decomposition pool structure.

	Biome-BGC	CLM-CN		
	$k_{disc1}(\text{d}^{-1})$	$k_{disc2} (\text{hr}^{-1})$	C:N ratio	Acceleration term (a_i)
k_{Lit1}	0.7	0.04892	•	1
k_{Lit2}	0.07	0.00302	•	1
k_{Lit3}	0.014	0.00059	•	1
k_{SOM1}	0.07	0.00302	12	1
k_{SOM2}	0.014	0.00059	12	1
k_{SOM3}	0.0014	0.00006	10	5
k_{SOM4}	0.0001	0.000004	10	70
k_{CWD}	0.001	0.00004	•	1

The first column of Table 2.27 gives the rates as used for the Biome-BGC model, which uses a discrete-time model with a daily timestep. The second column of Table 2.27 shows the rates transformed for a one-hour discrete timestep typical of CLM-CN. The transformation is based on the conversion of the initial discrete-time value (k_{disc1}) first to a continuous time value (k_{cont}), then to the new discrete-time value with a different timestep (k_{disc2}), following Olson (1963):

$$k_{cont} = -\log(1 - k_{disc1}) \quad (2.824)$$

$$k_{disc2} = 1 - \exp\left(-k_{cont} \frac{\Delta t_2}{\Delta t_1}\right) \quad (2.825)$$

where Δt_1 (s) and Δt_2 (s) are the time steps of the initial and new discrete-time models, respectively.

Respiration fractions are parameterized for decomposition fluxes out of each litter and SOM pool. The respiration fraction (rf , unitless) is the fraction of the decomposition carbon flux leaving one of the litter or SOM pools that is released as CO_2 due to heterotrophic respiration. Respiration fractions and exponential decay rates are estimated simultaneously from the results of microcosm decomposition experiments (Thornton, 1998). The same values are used in CLM-CN and Biome-BGC (Table 2.28).

Table 2.28: Respiration fractions for litter and SOM pools

Pool	rf
rf_{Lit1}	0.39
rf_{Lit2}	0.55
rf_{Lit3}	0.29
rf_{SOM1}	0.28
rf_{SOM2}	0.46
rf_{SOM3}	0.55
rf_{SOM4}	1.0 ^a

^aa The respiration fraction for pool SOM4 is 1.0 by definition: since there is no pool downstream of SOM4, the entire carbon flux leaving this pool is assumed to be respired as CO₂.

2.21.2 Century-based Pool Structure, Rate Constants and Parameters

The Century-based decomposition cascade is, like CLM-CN, a first-order decay model; the two structures differ in the number of pools, the connections between those pools, the turnover times of the pools, and the respiration fraction during each transition (Figure 15.2). The turnover times are different for the Century-based pool structure, following those described in Parton et al. (1988) (Table 2.29).

Table 2.29: Turnover times, C:N ratios, and acceleration parameters (see section 15.8 for explanation) for the Century-based decomposition cascade.

	Turnover time (year)	C:N ratio	Acceleration term (a_i)
CWD	4.1	•	1
Litter 1	0.066	•	1
Litter 2	0.25	•	1
Litter 3	0.25	•	1
SOM 1	0.17	8	1
SOM 2	6.1	11	15
SOM 3	270	11	675

Likewise, values for the respiration fraction of Century-based structure are in Table 2.30.

Table 2.30: Respiration fractions for litter and SOM pools for Century-based structure

Pool	rf
rf_{Lit1}	0.55
rf_{Lit2}	0.5
rf_{Lit3}	0.5
rf_{SOM1}	f(txt)
rf_{SOM2}	0.55
rf_{SOM3}	0.55

2.21.3 Environmental modifiers on decomposition rate

These base rates are modified on each timestep by functions of the current soil environment. For the single-level model, there are two rate modifiers, temperature (r_{tsoil} , unitless) and moisture (r_{water} , unitless), both of which are calculated using the average environmental conditions of the top five model levels (top 29 cm of soil column). For the vertically-resolved model, two additional environmental modifiers are calculated beyond the temperature and moisture limitations: an oxygen scalar (r_{oxygen} , unitless), and a depth scalar (r_{depth} , unitless).

The Temperature scalar r_{tsoil} is calculated in CLM using a Q_{10} approach, with $Q_{10} = 1.5$.

$$r_{tsoil} = Q_{10}^{\left(\frac{T_{soil,j} - T_{ref}}{10}\right)} \quad (2.826)$$

where j is the soil layer index, $T_{soil,j}$ (K) is the temperature of soil level j . The reference temperature $T_{ref} = 25\text{C}$.

The rate scalar for soil water potential (r_{water} , unitless) is calculated using a relationship from Andrén and Paustian (1987) and supported by additional data in Orchard and Cook (1983):

$$r_{water} = \sum_{j=1}^5 \left\{ \begin{array}{ll} 0 & \text{for } \Psi_j < \Psi_{min} \\ \frac{\log(\Psi_{min}/\Psi_j)}{\log(\Psi_{min}/\Psi_{max})} w_{soil,j} & \text{for } \Psi_{min} \leq \Psi_j \leq \Psi_{max} \\ 1 & \text{for } \Psi_j > \Psi_{max} \end{array} \right\} \quad (2.827)$$

where Ψ_j is the soil water potential in layer j , Ψ_{min} is a lower limit for soil water potential control on decomposition rate (in CLM5, this was changed from a default value of -10 MPa used in CLM4.5 and earlier to a default value of -2.5 MPa). $\Psi_{max,j}$ (MPa) is the soil moisture at which decomposition proceeds at a moisture-unlimited rate. The default value of $\Psi_{max,j}$ for CLM5 is updated from a saturated value used in CLM4.5 and earlier, to a value nominally at field capacity, with a value of -0.002 MPa

For frozen soils, the bulk of the rapid dropoff in decomposition with decreasing temperature is due to the moisture limitation, since matric potential is limited by temperature in the supercooled water formulation of Niu and Yang (2006),

$$\psi(T) = -\frac{L_f (T - T_f)}{10^3 T} \quad (2.828)$$

An additional frozen decomposition limitation can be specified using a ‘frozen Q_{10} ’ following Koven et al. (2011), however the default value of this is the same as the unfrozen Q_{10} value, and therefore the basic hypothesis is that frozen respiration is limited by liquid water availability, and can be modeled following the same approach as thawed but dry soils.

An additional rate scalar, r_{oxygen} is enabled when the CH₄ submodel is used (set equal to 1 for the single layer model or when the CH₄ submodel is disabled). This limits decomposition when there is insufficient molecular oxygen to satisfy stoichiometric demand (1 mol O₂ consumed per mol CO₂ produced) from heterotrophic decomposers, and supply from diffusion through soil layers (unsaturated and saturated) or aerenchyma (Chapter 19). A minimum value of r_{oxygen} is set at 0.2, with the assumption that oxygen within organic tissues can supply the necessary stoichiometric demand at this rate. This value lies between estimates of 0.025–0.1 (Frolking et al. 2001), and 0.35 (Wania et al. 2009); the large range of these estimates poses a large unresolved uncertainty.

Lastly, a possible explicit depth dependence, r_{depth} , (set equal to 1 for the single layer model) can be applied to soil C decomposition rates to account for processes other than temperature, moisture, and anoxia that can limit decomposition. This depth dependence of decomposition was shown by Jenkinson and Coleman (2008) to be an important term in fitting total C and 14C profiles, and implies that unresolved processes, such as priming effects, microscale anoxia, soil mineral surface and/or aggregate stabilization may be important in controlling the fate of carbon at depth Koven et al. (2013). CLM includes these unresolved depth controls via an exponential decrease in the soil turnover time with depth:

$$r_{depth} = \exp\left(-\frac{z}{z_\tau}\right) \quad (2.829)$$

where z_τ is the e-folding depth for decomposition. For CLM4.5, the default value of this was 0.5m. For CLM5, this has been changed to a default value of 10m, which effectively means that intrinsic decomposition rates may proceed as quickly at depth as at the surface.

The combined decomposition rate scalar (r_{total} , unitless) is:

$$r_{total} = r_{soil} r_{water} r_{oxygen} r_{depth}. \quad (2.830)$$

2.21.4 N-limitation of Decomposition Fluxes

Decomposition rates can also be limited by the availability of mineral nitrogen, but calculation of this limitation depends on first estimating the potential rates of decomposition, assuming an unlimited mineral nitrogen supply. The general case is described here first, referring to a generic decomposition flux from an “upstream” pool (u) to a “downstream” pool (d), with an intervening loss due to respiration. The potential carbon flux out of the upstream pool ($CF_{pot,u}$, gC m⁻² s⁻¹) is:

$$CF_{pot,u} = CS_u k_u \quad (2.831)$$

where CS_u (gC m⁻²) is the initial mass in the upstream pool and k_u is the decay rate constant (s:^{sup:-1}) for the upstream pool, adjusted for temperature and moisture conditions. Depending on the C:N ratios of the upstream and downstream pools and the amount of carbon lost in the transformation due to respiration (the respiration fraction), the execution of this potential carbon flux can generate either a source or a sink of new mineral nitrogen ($NF_{pot,min,u \rightarrow d}$, gN m⁻² s⁻¹). The governing equation (Thornton and Rosenbloom, 2005) is:

$$NF_{pot,min,u \rightarrow d} = \frac{CF_{pot,u} \left(1 - rf_u - \frac{CN_d}{CN_u}\right)}{CN_d} \quad (2.832)$$

where rf_u is the respiration fraction for fluxes leaving the upstream pool, CN_u and CN_d are the C:N ratios for upstream and downstream pools, respectively. Negative values of $NF_{pot,min,u \rightarrow d}$ indicate that the decomposition flux results in a source of new mineral nitrogen, while positive values indicate that the potential decomposition flux results in a sink (demand) for mineral nitrogen.

Following from the general case, potential carbon fluxes leaving individual pools in the decomposition cascade, for the example of the CLM-CN pool structure, are given as:

$$CF_{pot,Lit1} = CS_{Lit1} k_{Lit1} r_{total} / \Delta t \quad (2.833)$$

$$CF_{pot,Lit2} = CS_{Lit2} k_{Lit2} r_{total} / \Delta t \quad (2.834)$$

$$CF_{pot,Lit3} = CS_{Lit3} k_{Lit3} r_{total} / \Delta t \quad (2.835)$$

$$CF_{pot,SOM1} = CS_{SOM1} k_{SOM1} r_{total} / \Delta t \quad (2.836)$$

$$CF_{pot,SOM2} = CS_{SOM2} k_{SOM2} r_{total} / \Delta t \quad (2.837)$$

$$CF_{pot,SOM3} = CS_{SOM3} k_{SOM3} r_{total} / \Delta t \quad (2.838)$$

$$CF_{pot, SOM4} = CS_{SOM4} k_{SOM4} r_{total} / \Delta t \quad (2.839)$$

where the factor ($1/\Delta t$) is included because the rate constant is calculated for the entire timestep (Eqs. and), but the convention is to express all fluxes on a per-second basis. Potential mineral nitrogen fluxes associated with these decomposition steps are, again for the example of the CLM-CN pool structure (the CENTURY structure will be similar but without the different terminal step):

$$NF_{pot_min, Lit1 \rightarrow SOM1} = CF_{pot, Lit1} \left(1 - rf_{Lit1} - \frac{CN_{SOM1}}{CN_{Lit1}} \right) / CN_{SOM1} \quad (2.840)$$

$$NF_{pot_min, Lit2 \rightarrow SOM2} = CF_{pot, Lit2} \left(1 - rf_{Lit2} - \frac{CN_{SOM2}}{CN_{Lit2}} \right) / CN_{SOM2} \quad (2.841)$$

$$NF_{pot_min, Lit3 \rightarrow SOM3} = CF_{pot, Lit3} \left(1 - rf_{Lit3} - \frac{CN_{SOM3}}{CN_{Lit3}} \right) / CN_{SOM3} \quad (2.842)$$

$$NF_{pot_min, SOM1 \rightarrow SOM2} = CF_{pot, SOM1} \left(1 - rf_{SOM1} - \frac{CN_{SOM2}}{CN_{SOM1}} \right) / CN_{SOM2} \quad (2.843)$$

$$NF_{pot_min, SOM2 \rightarrow SOM3} = CF_{pot, SOM2} \left(1 - rf_{SOM2} - \frac{CN_{SOM3}}{CN_{SOM2}} \right) / CN_{SOM3} \quad (2.844)$$

$$NF_{pot_min, SOM3 \rightarrow SOM4} = CF_{pot, SOM3} \left(1 - rf_{SOM3} - \frac{CN_{SOM4}}{CN_{SOM3}} \right) / CN_{SOM4} \quad (2.845)$$

$$NF_{pot_min, SOM4} = -CF_{pot, SOM4} / CN_{SOM4} \quad (2.846)$$

where the special form of Eq. arises because there is no SOM pool downstream of SOM4 in the converging cascade: all carbon fluxes leaving that pool are assumed to be in the form of respired CO₂, and all nitrogen fluxes leaving that pool are assumed to be sources of new mineral nitrogen.

Steps in the decomposition cascade that result in release of new mineral nitrogen (mineralization fluxes) are allowed to proceed at their potential rates, without modification for nitrogen availability. Steps that result in an uptake of mineral nitrogen (immobilization fluxes) are subject to rate limitation, depending on the availability of mineral nitrogen, the total immobilization demand, and the total demand for soil mineral nitrogen to support new plant growth. The potential mineral nitrogen fluxes from Eqs. - are evaluated, summing all the positive fluxes to generate the total potential nitrogen immobilization flux (NF_{immob_demand} , gN m⁻² s⁻¹), and summing absolute values of all the negative fluxes to generate the total nitrogen mineralization flux (NF_{gross_nmin} , gN m⁻² s⁻¹). Since NF_{griss_nmin} is a source of new mineral nitrogen to the soil mineral nitrogen pool it is not limited by the availability of soil mineral nitrogen, and is therefore an actual as opposed to a potential flux.

2.21.5 N Competition between plant uptake and soil immobilization fluxes

Once NF_{immob_demand} and NF_{nit_demand} for each layer j are known, the competition between plant and microbial nitrogen demand can be resolved. Mineral nitrogen in the soil pool (NS_{smmin} , gN m⁻²) at the beginning of the timestep is considered the available supply.

Here, the NF_{plant_demand} is the theoretical maximum demand for nitrogen by plants to meet the entire carbon uptake given an N cost of zero (and therefore represents the upper bound on N requirements). N uptake costs that are > 0 imply that the plant will take up less N than it demands, ultimately. However, given the heuristic nature of the N competition algorithm, this discrepancy is not explicitly resolved here.

The hypothetical plant nitrogen demand from the soil mineral pool is distributed between layers in proportion to the profile of available mineral N:

$$NF_{plant_demand,j} = NF_{plant_demand} NS_{sminn,j} / \sum_{j=1}^{nj} NS_{sminn,j} \quad (2.847)$$

Plants first compete for ammonia (NH_4). For each soil layer (j), we calculate the total NH_4 demand as:

$$NF_{total_demand,nh4,j} = NF_{immob_demand,j} + NF_{immob_demand,j} + NF_{nit_demand,j} \quad (2.848)$$

where If $NF_{total_demand,j} \Delta t < NS_{sminn,j}$, then the available pool is large enough to meet both the maximum plant and microbial demand, then immobilization proceeds at the maximum rate.

$$f_{immob_demand,j} = 1.0 \quad (2.849)$$

where $f_{immob_demand,j}$ is the fraction of potential immobilization demand that can be met given current supply of mineral nitrogen in this layer. We also set the actual nitrification flux to be the same as the potential flux ($NF_{nit} = NF_{nit_demand}$).

If $NF_{total_demand,j} \Delta t \geq NS_{sminn,j}$, then there is not enough mineral nitrogen to meet the combined demands for plant growth and heterotrophic immobilization, immobilization is reduced proportional to the discrepancy, by $f_{immob_demand,j}$, where

$$f_{immob_demand,j} = \frac{NS_{sminn,j}}{\Delta t NF_{total_demand,j}} \quad (2.850)$$

The N available to the FUN model for plant uptake ($NF_{plant_avail_sminn}$ (gN m^{-2})), which determines both the cost of N uptake, and the absolute limit on the N which is available for acquisition, is calculated as the total mineralized pool minus the actual immobilized flux:

$$NF_{plant_avail_sminn,j} = NS_{sminn,j} - f_{immob_demand} NF_{immob_demand,j} \quad (2.851)$$

This treatment of competition for nitrogen as a limiting resource is referred to a demand-based competition, where the fraction of the available resource that eventually flows to a particular process depends on the demand from that process in comparison to the total demand from all processes. Processes expressing a greater demand acquire a larger vfraction of the available resource.

2.21.6 Final Decomposition Fluxes

With f_{immob_demand} known, final decomposition fluxes can be calculated. Actual carbon fluxes leaving the individual litter and SOM pools, again for the example of the CLM-CN pool structure (the CENTURY structure will be similar but, again without the different terminal step), are calculated as:

$$CF_{Lit1} = \left\{ \begin{array}{ll} CF_{pot, Lit1} f_{immob_demand} & \text{for } NF_{pot_min, Lit1 \rightarrow SOM1} > 0 \\ CF_{pot, Lit1} & \text{for } NF_{pot_min, Lit1 \rightarrow SOM1} \leq 0 \end{array} \right\} \quad (2.852)$$

$$CF_{Lit2} = \left\{ \begin{array}{ll} CF_{pot, Lit2} f_{immob_demand} & \text{for } NF_{pot_min, Lit2 \rightarrow SOM2} > 0 \\ CF_{pot, Lit2} & \text{for } NF_{pot_min, Lit2 \rightarrow SOM2} \leq 0 \end{array} \right\} \quad (2.853)$$

$$CF_{Lit3} = \begin{cases} CF_{pot, Lit3} f_{immob_demand} & \text{for } NF_{pot_min, Lit3 \rightarrow SOM3} > 0 \\ CF_{pot, Lit3} & \text{for } NF_{pot_min, Lit3 \rightarrow SOM3} \leq 0 \end{cases} \quad (2.854)$$

$$CF_{SOM1} = \begin{cases} CF_{pot, SOM1} f_{immob_demand} & \text{for } NF_{pot_min, SOM1 \rightarrow SOM2} > 0 \\ CF_{pot, SOM1} & \text{for } NF_{pot_min, SOM1 \rightarrow SOM2} \leq 0 \end{cases} \quad (2.855)$$

$$CF_{SOM2} = \begin{cases} CF_{pot, SOM2} f_{immob_demand} & \text{for } NF_{pot_min, SOM2 \rightarrow SOM3} > 0 \\ CF_{pot, SOM2} & \text{for } NF_{pot_min, SOM2 \rightarrow SOM3} \leq 0 \end{cases} \quad (2.856)$$

$$CF_{SOM3} = \begin{cases} CF_{pot, SOM3} f_{immob_demand} & \text{for } NF_{pot_min, SOM3 \rightarrow SOM4} > 0 \\ CF_{pot, SOM3} & \text{for } NF_{pot_min, SOM3 \rightarrow SOM4} \leq 0 \end{cases} \quad (2.857)$$

$$CF_{SOM4} = CF_{pot, SOM4} \quad (2.858)$$

Heterotrophic respiration fluxes (losses of carbon as CO₂ to the atmosphere) are:

$$CF_{Lit1, HR} = CF_{Lit1} r f_{Lit1} \quad (2.859)$$

$$CF_{Lit2, HR} = CF_{Lit2} r f_{Lit2} \quad (2.860)$$

$$CF_{Lit3, HR} = CF_{Lit3} r f_{Lit3} \quad (2.861)$$

$$CF_{SOM1, HR} = CF_{SOM1} r f_{SOM1} \quad (2.862)$$

$$CF_{SOM2, HR} = CF_{SOM2} r f_{SOM2} \quad (2.863)$$

$$CF_{SOM3, HR} = CF_{SOM3} r f_{SOM3} \quad (2.864)$$

$$CF_{SOM4, HR} = CF_{SOM4} r f_{SOM4} \quad (2.865)$$

Transfers of carbon from upstream to downstream pools in the decomposition cascade are given as:

$$CF_{Lit1, SOM1} = CF_{Lit1} (1 - r f_{Lit1}) \quad (2.866)$$

$$CF_{Lit2, SOM2} = CF_{Lit2} (1 - r f_{Lit2}) \quad (2.867)$$

$$CF_{Lit3, SOM3} = CF_{Lit3} (1 - rf_{Lit3}) \quad (2.868)$$

$$CF_{SOM1, SOM2} = CF_{SOM1} (1 - rf_{SOM1}) \quad (2.869)$$

$$CF_{SOM2, SOM3} = CF_{SOM2} (1 - rf_{SOM2}) \quad (2.870)$$

$$CF_{SOM3, SOM4} = CF_{SOM3} (1 - rf_{SOM3}) \quad (2.871)$$

In accounting for the fluxes of nitrogen between pools in the decomposition cascade and associated fluxes to or from the soil mineral nitrogen pool, the model first calculates a flux of nitrogen from an upstream pool to a downstream pool, then calculates a flux either from the soil mineral nitrogen pool to the downstream pool (immobilization) or from the downstream pool to the soil mineral nitrogen pool (mineralization). Transfers of nitrogen from upstream to downstream pools in the decomposition cascade are given as:

$$NF_{Lit1, SOM1} = CF_{Lit1}/CN_{Lit1} \quad (2.872)$$

$$NF_{Lit2, SOM2} = CF_{Lit2}/CN_{Lit2} \quad (2.873)$$

$$NF_{Lit3, SOM3} = CF_{Lit3}/CN_{Lit3} \quad (2.874)$$

$$NF_{SOM1, SOM2} = CF_{SOM1}/CN_{SOM1} \quad (2.875)$$

$$NF_{SOM2, SOM3} = CF_{SOM2}/CN_{SOM2} \quad (2.876)$$

$$NF_{SOM3, SOM4} = CF_{SOM3}/CN_{SOM3} \quad (2.877)$$

Corresponding fluxes to or from the soil mineral nitrogen pool depend on whether the decomposition step is an immobilization flux or a mineralization flux:

$$NF_{smminn, Lit1 \rightarrow SOM1} = \begin{cases} NF_{pot_min, Lit1 \rightarrow SOM1} f_{immob_demand} & \text{for } NF_{pot_min, Lit1 \rightarrow SOM1} > 0 \\ NF_{pot_min, Lit1 \rightarrow SOM1} & \text{for } NF_{pot_min, Lit1 \rightarrow SOM1} \leq 0 \end{cases} \quad (2.878)$$

$$NF_{smminn, Lit2 \rightarrow SOM2} = \begin{cases} NF_{pot_min, Lit2 \rightarrow SOM2} f_{immob_demand} & \text{for } NF_{pot_min, Lit2 \rightarrow SOM2} > 0 \\ NF_{pot_min, Lit2 \rightarrow SOM2} & \text{for } NF_{pot_min, Lit2 \rightarrow SOM2} \leq 0 \end{cases} \quad (2.879)$$

$$NF_{smminn, Lit3 \rightarrow SOM3} = \begin{cases} NF_{pot_min, Lit3 \rightarrow SOM3} f_{immob_demand} & \text{for } NF_{pot_min, Lit3 \rightarrow SOM3} > 0 \\ NF_{pot_min, Lit3 \rightarrow SOM3} & \text{for } NF_{pot_min, Lit3 \rightarrow SOM3} \leq 0 \end{cases} \quad (2.880)$$

$$NF_{smminn, SOM1 \rightarrow SOM2} = \begin{cases} NF_{pot_min, SOM1 \rightarrow SOM2} f_{immob_demand} & \text{for } NF_{pot_min, SOM1 \rightarrow SOM2} > 0 \\ NF_{pot_min, SOM1 \rightarrow SOM2} & \text{for } NF_{pot_min, SOM1 \rightarrow SOM2} \leq 0 \end{cases} \quad (2.881)$$

$$NF_{smminn, SOM2 \rightarrow SOM3} = \begin{cases} NF_{pot_min, SOM2 \rightarrow SOM3} f_{immob_demand} & \text{for } NF_{pot_min, SOM2 \rightarrow SOM3} > 0 \\ NF_{pot_min, SOM2 \rightarrow SOM3} & \text{for } NF_{pot_min, SOM2 \rightarrow SOM3} \leq 0 \end{cases} \quad (2.882)$$

$$NF_{smminn, SOM3 \rightarrow SOM4} = \begin{cases} NF_{pot_min, SOM3 \rightarrow SOM4} f_{immob_demand} & \text{for } NF_{pot_min, SOM3 \rightarrow SOM4} > 0 \\ NF_{pot_min, SOM3 \rightarrow SOM4} & \text{for } NF_{pot_min, SOM3 \rightarrow SOM4} \leq 0 \end{cases} \quad (2.883)$$

$$NF_{smminn, SOM4} = NF_{pot_min, SOM4} \quad (2.884)$$

2.21.7 Vertical Distribution and Transport of Decomposing C and N pools

Additional terms are needed to calculate the vertically-resolved soil C and N budget: the initial vertical distribution of C and N from PFTs delivered to the litter and CWD pools, and the vertical transport of C and N pools.

For initial vertical inputs, CLM uses separate profiles for aboveground (leaf, stem) and belowground (root) inputs. Aboveground inputs are given a single exponential with default e-folding depth = 0.1m. Belowground inputs are distributed according to rooting profiles with default values based on the Jackson et al. (1996) exponential parameterization.

Vertical mixing is accomplished by an advection-diffusion equation. The goal of this is to consider slow, soil- and adsorbed-phase transport due to bioturbation, cryoturbation, and erosion. Faster aqueous-phase transport is not included in CLM, but has been developed as part of the CLM-BeTR suite of parameterizations (Tang and Riley 2013). The default value of the advection term is 0 cm/yr, such that transport is purely diffusive. Diffusive transport differs in rate between permafrost soils (where cryoturbation is the dominant transport term) and non-permafrost soils (where bioturbation dominates). For permafrost soils, a parameterization based on that of Koven et al. (2009) is used: the diffusivity parameter is constant through the active layer, and decreases linearly from the base of the active layer to zero at a set depth (default 3m); the default permafrost diffusivity is 5 cm²/yr. For non-permafrost soils, the default diffusivity is 1 cm²/yr.

2.21.8 Model Equilibration and its Acceleration

For transient experiments, it is usually assumed that the carbon cycle is starting from a point of relatively close equilibrium, i.e. that productivity is balanced by ecosystem carbon losses through respiratory and disturbance pathways. In order to satisfy this assumption, the model is generally run until the productivity and loss terms find a stable long-term equilibrium; at this point the model is considered ‘spun up’.

Because of the coupling between the slowest SOM pools and productivity through N downregulation of photosynthesis, equilibration of the model for initialization purposes will take an extremely long time in the standard mode. This is particularly true for the CENTURY-based decomposition cascade, which includes a passive pool. In order to rapidly equilibrate the model, a modified version of the “accelerated decomposition” (Thornton and Rosenbloom, 2005) is

used. The fundamental idea of this approach is to allow fluxes between the various pools (both turnover-defined and vertically-defined fluxes) adjust rapidly, while keeping the pool sizes themselves small so that they can fill quickly. To do this, the base decomposition rate k_i for each pool i is accelerated by a term a_i such that the slow pools are collapsed onto an approximately annual timescale [Koven et al. \(2013\)](#). Accelerating the pools beyond this timescale distorts the seasonal and/or diurnal cycles of decomposition and N mineralization, thus leading to a substantially different ecosystem productivity than the full model. For the vertical model, the vertical transport terms are also accelerated by the same term a_i , as is the radioactive decay when ^{14}C is enabled, following the same principle of keeping fluxes between pools (or fluxes lost to decay) close to the full model while keeping the pools sizes small. When leaving the accelerated decomposition mode, the concentration of C and N in pools that had been accelerated are multiplied by the same term a_i , to bring the model into approximate equilibrium. Note that in CLM, the model can also transition into accelerated decomposition mode from the standard mode (by dividing the pools by a_i), and that the transitions into and out of accelerated decomposition mode are handled automatically by CLM upon loading from restart files (which preserve information about the mode of the model when restart files were written).

The base acceleration terms for the two decomposition cascades are shown in Tables 15.1 and 15.3. In addition to the base terms, CLM5 also includes a geographic term to the acceleration in order to apply larger values to high-latitude systems, where decomposition rates are particularly slow and thus equilibration can take significantly longer than in temperate or tropical climates. This geographic term takes the form of a logistic equation, where a_l is equal to the product of the base acceleration term and a_l below:

$$a_l = 1 + 50 / (1 + \exp(-0.1 * (\text{abs}(\text{latitude}) - 60))) \quad (2.885)$$

2.22 External Nitrogen Cycle

2.22.1 Summary of CLM5.0 updates relative to CLM4.5

We describe external inputs to the nitrogen cycle in CLM5.0. Much of the following information appeared in the CLM4.5 Technical Note ([Oleson et al. 2013](#)) as well as [Koven et al. \(2013\)](#).

CLM5.0 includes the following changes to terrestrial nitrogen inputs:

- Time varying deposition of reactive nitrogen. In off-line runs this changes monthly. In coupled simulations N deposition is passed at the coupling timestep (e.g., half-hourly).
- Asymbiotic (or free living) N fixation is a function of evapotranspiration and is added to the inorganic nitrogen (NH_4^+) pool (described below).
- Symbiotic N fixation is handled by the FUN model (chapter 2.18) and is passed straight to the plant, not the mineral nitrogen pool.

2.22.2 Overview

In addition to the relatively rapid cycling of nitrogen within the plant – litter – soil organic matter system, CLM also represents several processes which couple the internal nitrogen cycle to external sources and sinks. Inputs of new mineral nitrogen are from atmospheric deposition and biological nitrogen fixation. Losses of mineral nitrogen are due to nitrification, denitrification, leaching, and losses in fire. While the short-term dynamics of nitrogen limitation depend on the behavior of the internal nitrogen cycle, establishment of total ecosystem nitrogen stocks depends on the balance between sources and sinks in the external nitrogen cycle ([Thomas et al. 2015](#)).

As with CLM4.5, CLM5.0 represents inorganic N transformations based on the Century N-gas model; this includes separate NH_4^+ and NO_3^- pools, as well as environmentally controlled nitrification and denitrification rates that is described below.

2.22.3 Atmospheric Nitrogen Deposition

CLM uses a single variable to represent the total deposition of mineral nitrogen onto the land surface, combining wet and dry deposition of NO_y and NH_x as a single flux (NF_{ndep_sminn} , $\text{gN m}^{-2} \text{ s}^{-1}$). This flux is intended to represent total reactive nitrogen deposited to the land surface which originates from the following natural and anthropogenic sources (Galloway et al. 2004): formation of NO_x during lightning, NO_x and NH_3 emission from wildfire, NO_x emission from natural soils, NH_3 emission from natural soils, vegetation, and wild animals, NO_x and NH_3 emission during fossil fuel combustion (both thermal and fuel NO_x production), NO_x and NH_3 emission from other industrial processes, NO_x and NH_3 emission from fire associated with deforestation, NO_x and NH_3 emission from agricultural burning, NO_x emission from agricultural soils, NH_3 emission from agricultural crops, NH_3 emission from agricultural animal waste, and NH_3 emission from human waste and waste water. The deposition flux is provided as a spatially and (potentially) temporally varying dataset (see section 2.2.3 for a description of the default input dataset).

The nitrogen deposition flux is assumed to enter the NH_4^+ pool, and is vertically distributed throughout the soil profile. Although N deposition inputs include both oxidized and reduced forms, CLM5 only reads in total N deposition. This approach is held over from CLM4.0, which only represented a single mineral nitrogen pool, however, real pathways for wet and dry nitrogen deposition can be more complex than currently represented in the CLM5.0, including release from melting snowpack and direct foliar uptake of deposited NO_y (Tye et al. 2005; Vallano and Sparks, 2007).

In offline (uncoupled) CLM5.0 simulations monthly estimates of N deposition are provided, as opposed to decadal files supplied with previous versions of the model. In coupled simulations, N depositions fluxes are passed to the land model at the frequency of the time step (every half hour) through the coupler.

2.22.4 Biological Nitrogen Fixation

The fixation of new reactive nitrogen from atmospheric N_2 by soil microorganisms is an important component of both preindustrial and modern-day nitrogen budgets, but a mechanistic understanding of global-scale controls on biological nitrogen fixation (BNF) is still only poorly developed (Cleveland et al. 1999; Galloway et al. 2004). CLM5.0 uses the FUN model (chapter 2.18) to calculate the carbon cost and nitrogen acquired through symbiotic nitrogen fixation. This nitrogen is immediately available to plants.

Cleveland et al. (1999) suggested an empirical relationships that predicts BNF as a function of either evapotranspiration rate or net primary productivity for natural vegetation. CLM5.0 adopts the evapotranspiration approach to calculate asymbiotic, or free-living, N fixation. This function has been modified from the Cleveland et al. (1999) estimates to provide lower estimate of free-living nitrogen fixation in CLM5.0 (CF_{ann_ET} , mm yr^{-1}). This moves away from the NPP approach used in CLM4.0 and 4.5 and avoids unrealistically increasing freeliving rates of N fixation under global change scenarios (Wieder et al. 2015). The expression used is:

$$NF_{nfix,sminn} = 0.0006 (0.0117 + CF_{ann_ET}) / (86400 \cdot 365) \quad (2.886)$$

Where $NF_{nfix,sminn}$ ($\text{gN m}^{-2} \text{ s}^{-1}$) is the rate of free-living nitrogen fixation in Figure 2.18.

As with Atmospheric N deposition, free-living N inputs are added directly to the NH_4^+ pool.

2.22.5 Nitrification and Denitrification Losses of Nitrogen

Nitrification is an autotrophic process that converts less mobile ammonium ions into nitrate, that can more easily be lost from soil systems by leaching or denitrification. The process catalyzed by ammonia oxidizing archaea and bacteria that convert ammonium (NH_4^+) into nitrite, which is subsequently oxidized into nitrate (NO_3^-). Conditions favoring nitrification include high NH_4^+ concentrations, well aerated soils, a neutral pH and warmer temperatures.

Under aerobic conditions in the soil oxygen is the preferred electron acceptor supporting the metabolism of heterotrophs, but anaerobic conditions favor the activity of soil heterotrophs which use nitrate as an electron acceptor (e.g. *Pseudomonas* and *Clostridium*) supporting respiration. This process, known as denitrification, results in the transformation of nitrate to gaseous N_2 , with smaller associated production of NO_x and N_2O . It is typically assumed

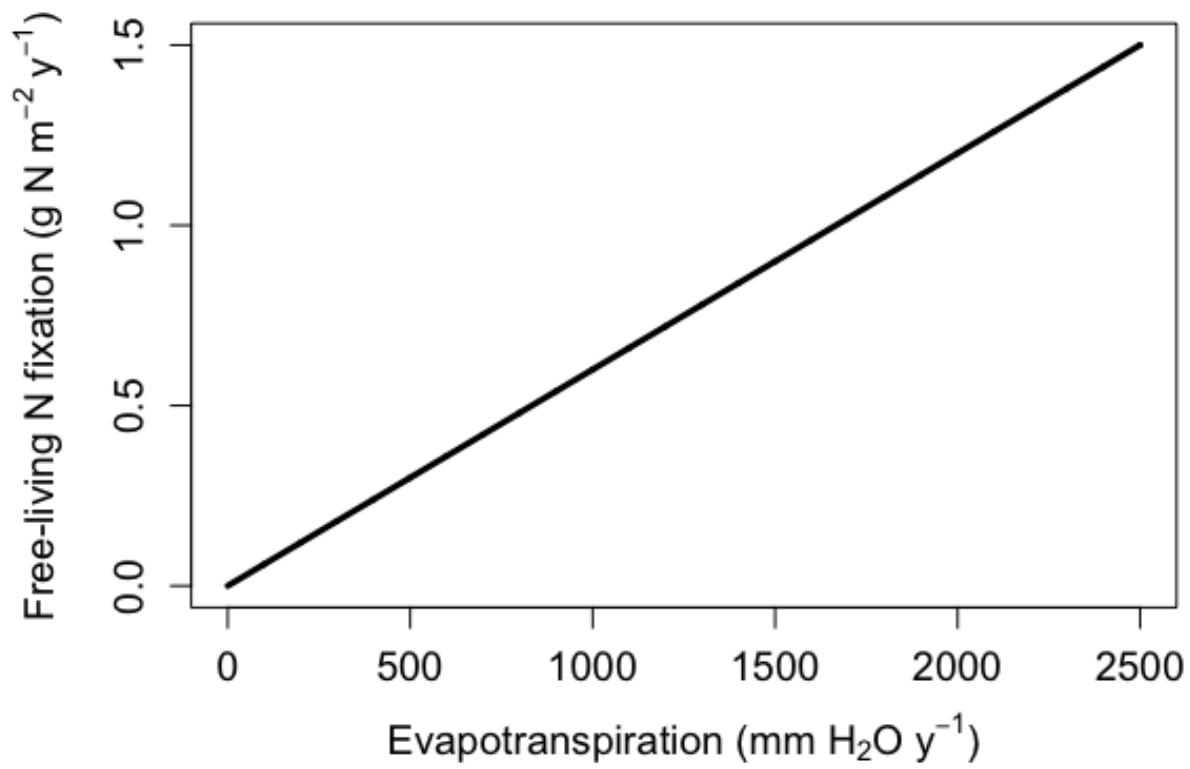


Figure 2.18: Free-living nitrogen fixation as a function of annual evapotranspiration. Results here show annual N inputs from free-living N fixations, but the model actually calculates inputs on a per second basis.

that nitrogen fixation and denitrification were approximately balanced in the preindustrial biosphere (*Galloway et al. 2004*). It is likely that denitrification can occur within anaerobic microsites within an otherwise aerobic soil environment, leading to large global denitrification fluxes even when fluxes per unit area are rather low (*Galloway et al. 2004*).

CLM includes a detailed representation of nitrification and denitrification based on the Century N model (*Parton et al. 1996, 2001; del Grosso et al. 2000*). In this approach, nitrification of NH_4^+ to NO_3^- is a function of temperature, moisture, and pH:

$$f_{nitr,p} = [\text{NH}_4] k_{nitr} f(T) f(H_2O) f(pH) \quad (2.887)$$

where $f_{nitr,p}$ is the potential nitrification rate (prior to competition for NH_4^+ by plant uptake and N immobilization), k_{nitr} is the maximum nitrification rate (10 % day $^{-1}$, (*Parton et al. 2001*)), and $f(T)$ and $f(H_2O)$ are rate modifiers for temperature and moisture content. CLM uses the same rate modifiers as are used in the decomposition routine. $f(pH)$ is a rate modifier for pH; however, because CLM does not calculate pH, instead a fixed pH value of 6.5 is used in the pH function of *Parton et al. (1996)*.

The potential denitrification rate is co-limited by NO_3^- concentration and C consumption rates, and occurs only in the anoxic fraction of soils:

$$f_{denitr,p} = \min(f(decomp), f([NO_3^-])) frac_{anox} \quad (2.888)$$

where $f_{denitr,p}$ is the potential denitrification rate and $f(decomp)$ and $f([NO_3^-])$ are the carbon- and nitrate- limited denitrification rate functions, respectively, (*del Grosso et al. 2000*). Because the modified CLM includes explicit treatment of soil biogeochemical vertical profiles, including diffusion of the trace gases O_2 and CH_4 (*Riley et al. 2011a*), the calculation of anoxic fraction $frac_{anox}$ uses this information following the anoxic microsite formulation of *Arah and Vinten (1995)*.

$$frac_{anox} = \exp(-aR_\psi^{-\alpha}V^{-\beta}C^\gamma[\theta + \chi\varepsilon]^\delta) \quad (2.889)$$

where a , α , β , γ , and δ are constants (equal to 1.5×10^{-10} , 1.26, 0.6, 0.6, and 0.85, respectively), R_ψ is the radius of a typical pore space at moisture content ψ , V is the O_2 consumption rate, C is the O_2 concentration, θ is the water-filled pore space, χ is the ratio of diffusivity of oxygen in water to that in air, and ε is the air-filled pore space (*Arah and Vinten (1995)*). These parameters are all calculated separately at each layer to define a profile of anoxic porespace fraction in the soil.

The nitrification/denitrification models used here also predict fluxes of N_2O via a “hole-in-the-pipe” approach (*Firestone and Davidson, 1989*). A constant fraction ($6 * 10^{-4}$, *Li et al. 2000*) of the nitrification flux is assumed to be N_2O , while the fraction of denitrification going to N_2O , $P_{N2:N2O}$, is variable, following the Century (*del Grosso et al. 2000*) approach:

$$P_{N2:N2O} = \max(0.16k_1, k_1 \exp(-0.8P_{NO_3:CO_2})) f_{WFPS} \quad (2.890)$$

where $P_{NO_3:CO_2}$ is the ratio of CO_2 production in a given soil layer to the NO_3^- concentration, k_1 is a function of d_g , the gas diffusivity through the soil matrix:

$$k_1 = \max(1.7, 38.4 - 350 * d_g) \quad (2.891)$$

and f_{WFPS} is a function of the water filled pore space $WFPS$:

$$f_{WFPS} = \max(0.1, 0.015 \times WFPS - 0.32) \quad (2.892)$$

2.22.6 Leaching Losses of Nitrogen

Soil mineral nitrogen remaining after plant uptake, immobilization, and denitrification is subject to loss as a dissolved component of hydrologic outflow from the soil column (leaching). This leaching loss ($NF_{leached}$, $\text{gN m}^{-2} \text{s}^{-1}$) depends

on the concentration of dissolved mineral (inorganic) nitrogen in soil water solution (DIN , gN kg H_2O), and the rate of hydrologic discharge from the soil column to streamflow (Q_{dis} , kg H_2O m $^{-2}$ s $^{-1}$, section 2.7.5), as

$$NF_{leached} = DIN \cdot Q_{dis}. \quad (2.893)$$

DIN is calculated assuming that a constant fraction (sf , proportion) of the remaining soil mineral N pool is in soluble form, and that this entire fraction is dissolved in the total soil water. For the Century- based formulation in CLM5.0, the leaching acts only on the NO_3^- pool (which is assumed to be 100% soluble), while the NH_4^+ pool is assumed to be 100% adsorbed onto mineral surfaces and unaffected by leaching. DIN is then given as

$$DIN = \frac{NS_{sminn}sf}{WS_{tot_soil}} \quad (2.894)$$

where WS_{tot_soil} (kgH:sub:2O m $^{-2}$) is the total mass of soil water content integrated over the column. The total mineral nitrogen leaching flux is limited on each time step to not exceed the soluble fraction of NS_{sminn}

$$NF_{leached} = \min \left(NF_{leached}, \frac{NS_{sminn}sf}{\Delta t} \right). \quad (2.895)$$

2.22.7 Losses of Nitrogen Due to Fire

The final pathway for nitrogen loss is through combustion, also known as pyrodenitrification. Detailed equations are provided, together with the effects of fire on the carbon budget, in Chapter 2.23). It is assumed in CLM-CN that losses of N due to fire are restricted to vegetation and litter pools (including coarse woody debris). Loss rates of N are determined by the fraction of biomass lost to combustion, assuming that most of the nitrogen in the burned biomass is lost to the atmosphere (Schlesinger, 1997; Smith et al. 2005). It is assumed that soil organic matter pools of carbon and nitrogen are not directly affected by fire (Neff et al. 2005).

2.23 Fire

The fire parameterization in CLM contains four components: non-peat fires outside cropland and tropical closed forests, agricultural fires in cropland, deforestation fires in the tropical closed forests, and peat fires (see Li et al. 2012a, Li et al. 2012b, Li et al. 2013, Li and Lawrence 2017 for details). In this fire parameterization, burned area is affected by climate and weather conditions, vegetation composition and structure, and human activities. After burned area is calculated, we estimate the fire impact, including biomass and peat burning, fire-induced vegetation mortality, adjustment of the carbon and nitrogen (C/N) pools, and fire emissions.

2.23.1 Non-peat fires outside cropland and tropical closed forest

Burned area in a grid cell, A_b (km 2 s $^{-1}$), is determined by

$$A_b = N_f a \quad (2.896)$$

where N_f (count s $^{-1}$) is fire counts in the grid cell; a (km 2) is average fire spread area of a fire.

Fire counts

Fire counts N_f is taken as

$$N_f = N_i f_b f_m f_{se,o} \quad (2.897)$$

where N_i (count s^{-1}) is the number of ignition sources due to natural causes and human activities; f_b and f_m (fractions) represent the availability and combustibility of fuel, respectively; $f_{se,o}$ is the fraction of anthropogenic and natural fires unsuppressed by humans and related to the socioeconomic conditions.

N_i (count s^{-1}) is given as

$$N_i = (I_n + I_a) A_g \quad (2.898)$$

where I_n (count $km^{-2} s^{-1}$) and I_a (count $km^{-2} s^{-1}$) are the number of natural and anthropogenic ignitions per km^2 , respectively; A_g is the area of the grid cell (km^2). I_n is estimated by

$$I_n = \gamma \psi I_l \quad (2.899)$$

where $\gamma = 0.22$ is ignition efficiency of cloud-to-ground lightning; $\psi = \frac{1}{5.16 + 2.16 \cos[3\min(60, \lambda)]}$ is the cloud-to-ground lightning fraction and depends on the latitude λ (degrees); I_l (flash $km^{-2} s^{-1}$) is the total lightning flashes. I_a is modeled as a monotonic increasing function of population density:

$$I_a = \frac{\alpha D_P k(D_P)}{n} \quad (2.900)$$

where $\alpha = 0.01$ (count $person^{-1} mon^{-1}$) is the number of potential ignition sources by a person per month; D_P (person km^{-2}) is the population density; $k(D_P) = 6.8D_P^{-0.6}$ represents anthropogenic ignition potential as a function of human population density D_P ; n is the seconds in a month.

Fuel availability f_b is given as

$$f_b = \begin{cases} 0 & B_{ag} < B_{low} \\ \frac{B_{ag} - B_{low}}{B_{up} - B_{low}} & B_{low} \leq B_{ag} \leq B_{up} \\ 1 & B_{ag} > B_{up} \end{cases}, \quad (2.901)$$

where B_{ag} ($g C m^{-2}$) is the biomass of combined leaf, stem, litter, and woody debris pools; $B_{low} = 105 g C m^{-2}$ is the lower fuel threshold below which fire does not occur; $B_{up} = 1050 g C m^{-2}$ is the upper fuel threshold above which fire occurrence is not limited by fuel availability.

Fuel combustibility f_m is estimated by

$$f_m = f_{RH} f_\beta, \quad T_{17cm} > T_f \quad (2.902)$$

where f_{RH} and f_β represent the dependence of fuel combustibility on relative humidity RH (%) and root-zone soil moisture limitation β (fraction); T_{17cm} is the temperature of the top 17 cm of soil (K) and T_f is the freezing temperature. f_{RH} is a weighted average of real time RH (RH_0) and 30-day running mean RH (RH_{30d}):

$$f_{RH} = (1 - w)l_{RH_0} + wl_{RH_{30d}} \quad (2.903)$$

where weight $w = \max[0, \min(1, \frac{B_{ag} - 2500}{2500})]$, $l_{RH_0} = 1 - \max[0, \min(1, \frac{RH_0 - 30}{80 - 30})]$, and $l_{RH_{30d}} = 1 - \max[0.75, \min(1, \frac{RH_{30d}}{90})]$. f_β is given by

$$f_\beta = \begin{cases} 1 & \beta \leq \beta_{low} \\ \frac{\beta_{up} - \beta}{\beta_{up} - \beta_{low}} & \beta_{low} < \beta < \beta_{up} \\ 0 & \beta \geq \beta_{up} \end{cases}, \quad (2.904)$$

where $\beta_{low} = 0.85$ and $\beta_{up} = 0.98$ are the lower and upper thresholds, respectively.

For scarcely populated regions ($D_p \leq 0.1$ person km^{-2}), we assume that anthropogenic suppression on fire occurrence is negligible, i.e., $f_{se,o} = 1.0$. In regions of $D_p > 0.1$ person km^{-2} , we parameterize the fraction of anthropogenic and natural fires unsuppressed by human activities as

$$f_{se,o} = f_d f_e \quad (2.905)$$

where f_d and f_e are the effects of the demographic and economic conditions on fire occurrence. The demographic influence on fire occurrence is

$$f_d = 0.01 + 0.98 \exp(-0.025D_P). \quad (2.906)$$

For shrub and grass PFTs, the economic influence on fire occurrence is parameterized as a function of Gross Domestic Product GDP (k 1995US\$ capita⁻¹):

$$f_e = 0.1 + 0.9 \times \exp[-\pi(\frac{GDP}{8})^{0.5}] \quad (2.907)$$

which captures 73% of the observed MODIS fire counts with variable GDP in regions where shrub and grass PFTs are dominant (fractional coverage of shrub and grass PFTs > 50%). In regions outside tropical closed forests and dominated by trees (fractional coverage of tree PFTs > 50%), we use

$$f_e = \begin{cases} 0.39 & GDP > 20 \\ 0.79 & 8 < GDP \leq 20 \\ 1 & GDP \leq 8 \end{cases}, \quad (2.908)$$

to reproduce the relationship between MODIS fire counts and GDP.

Average spread area of a fire

Fire fighting capacity depends on socioeconomic conditions and affects fire spread area. Due to a lack of observations, we consider the socioeconomic impact on the average burned area rather than separately on fire spread rate and fire duration:

$$a = a^* F_{se} \quad (2.909)$$

where a^* is the average burned area of a fire without anthropogenic suppression and F_{se} is the socioeconomic effect on fire spread area.

Average burned area of a fire without anthropogenic suppression is assumed elliptical in shape with the wind direction along the major axis and the point of ignition at one of the foci. According to the area formula for an ellipse, average burned area of a fire can be represented as:

$$a^* = \pi \frac{l}{2} \frac{w}{2} \times 10^{-6} = \frac{\pi u_p^2 \tau^2}{4L_B} (1 + \frac{1}{H_B})^2 \times 10^{-6} \quad (2.910)$$

where u_p (m s⁻¹) is the fire spread rate in the downwind direction; τ (s) is average fire duration; L_B and H_B are length-to-breadth ratio and head-to-back ratio of the ellipse; 10⁻⁶ converts m² to km².

According to *Arora and Boer (2005)*,

$$L_B = 1.0 + 10.0[1 - \exp(-0.06W)] \quad (2.911)$$

where W (m s⁻¹) is the wind speed. According to the mathematical properties of the ellipse, the head-to-back ratio H_B is

$$H_B = \frac{u_p}{u_b} = \frac{L_B + (L_B^2 - 1)^{0.5}}{L_B - (L_B^2 - 1)^{0.5}}. \quad (2.912)$$

The fire spread rate in the downwind direction is represented as

$$u_p = u_{\max} C_m g(W) \quad (2.913)$$

(*Arora and Boer, 2005*), where u_{\max} (m s⁻¹) is the PFT-dependent average maximum fire spread rate in natural vegetation regions; $C_m = \sqrt{f_m}$ and $g(W)$ represent the dependence of u_p on fuel wetness and wind speed W , respectively.

u_{\max} is set to 0.33 m s^{-1} for grass PFTs, 0.28 m s^{-1} for shrub PFTs, 0.26 m s^{-1} for needleleaf tree PFTs, and 0.25 m s^{-1} for other tree PFTs. $g(W)$ is derived from the mathematical properties of the ellipse and equation (2.911) and (2.912).

$$g(W) = \frac{2L_B}{1 + \frac{1}{H_B}} g(0). \quad (2.914)$$

Since $g(W)=1.0$, and L_B and H_B are at their maxima $L_B^{\max} = 11.0$ and $H_B^{\max} = 482.0$ when $W \rightarrow \infty$, $g(0)$ can be derived as

$$g(0) = \frac{1 + \frac{1}{H_B^{\max}}}{2L_B^{\max}} = 0.05. \quad (2.915)$$

In the absence of globally gridded data on barriers to fire (e.g. rivers, lakes, roads, firebreaks) and human fire-fighting efforts, average fire duration is simply assumed equal to 1 which is the observed 2001–2004 mean persistence of most fires in the world (*Giglio et al. 2006*).

As with the socioeconomic influence on fire occurrence, we assume that the socioeconomic influence on fire spreading is negligible in regions of $D_p \leq 0.1 \text{ person km}^{-2}$, i.e., $F_{se} = 1.0$. In regions of $D_p > 0.1 \text{ person km}^{-2}$, we parameterize such socioeconomic influence as:

$$F_{se} = F_d F_e \quad (2.916)$$

where F_d and F_e are effects of the demographic and economic conditions on the average spread area of a fire, and are identified by maximizing the explained variability of the GFED3 burned area fraction with both socioeconomic indices in grid cells with various dominant vegetation types. For shrub and grass PFTs, the demographic impact factor is

$$F_d = 0.2 + 0.8 \times \exp[-\pi(\frac{D_p}{450})^{0.5}] \quad (2.917)$$

and the economic impact factor is

$$F_e = 0.2 + 0.8 \times \exp(-\pi \frac{GDP}{7}). \quad (2.918)$$

For tree PFTs outside tropical closed forests, the demographic and economic impact factors are given as

$$F_d = 0.4 + 0.6 \times \exp(-\pi \frac{D_p}{125}) \quad (2.919)$$

and

$$F_e = \begin{cases} 0.62, & GDP > 20 \\ 0.83, & 8 < GDP \leq 20 \\ 1, & GDP \leq 8 \end{cases}. \quad (2.920)$$

Equations (2.917) - (2.920) reflect that more developed and more densely populated regions have a higher fire fighting capability.

Fire impact

In post-fire regions, we calculate PFT-level fire carbon emissions from biomass burning of the j th PFT, ϕ_j (g C s^{-1}), as

$$\phi_j = A_{b,j} \mathbf{C}_j \bullet \mathbf{CC}_j \quad (2.921)$$

where $A_{b,j}$ ($\text{km}^2 \text{ s}^{-1}$) is burned area for the j th PFT; $\mathbf{C}_j = (C_{leaf}, C_{stem}, C_{root}, C_{ts})$ is a vector with carbon density (g C km^{-2}) for leaf, stem (live and dead stem), root (fine, live coarse and dead coarse root), and transfer and storage

carbon pools as elements; $\mathbf{CC}_j = (\mathbf{CC}_{leaf}, \mathbf{CC}_{stem}, \mathbf{CC}_{root}, \mathbf{CC}_{ts})$ is the corresponding combustion completeness factor vector (Table 2.31). Moreover, we assume that 50% and 28% of column-level litter and coarse woody debris are burned and the corresponding carbon is transferred to atmosphere.

Tissue mortality due to fire leads to carbon transfers in two ways. First, carbon from uncombusted leaf, live stem, dead stem, root, and transfer and storage pools $\mathbf{C}'_{j1} = (C_{leaf}(1 - CC_{leaf}), C_{livestem}(1 - CC_{stem}), C_{deadstem}(1 - CC_{stem}), C_{root}(1 - CC_{root}), C_{ts}(1 - CC_{ts}))_j$ (g C km^{-2}) is transferred to litter as

$$\Psi_{j1} = \frac{A_{b,j}}{f_j A_g} \mathbf{C}'_{j1} \bullet M_{j1} \quad (2.922)$$

where $M_{j1} = (M_{leaf}, M_{livestem,1}, M_{deadstem}, M_{root}, M_{ts})_j$ is the corresponding mortality factor vector (Table 2.31). Second, carbon from uncombusted live stems is transferred to dead stems as:

$$\Psi_{j2} = \frac{A_{b,j}}{f_j A_g} C_{livestem}(1 - CC_{stem}) M_{livestem,2} \quad (2.923)$$

where $M_{livestem,2}$ is the corresponding mortality factor (Table 2.31).

Fire nitrogen emissions and nitrogen transfers due to fire-induced mortality are calculated the same way as for carbon, using the same values for combustion completeness and mortality factors. With CLM's dynamic vegetation option enabled, the number of tree PFT individuals killed by fire per km^2 (individual $\text{km}^{-2} \text{s}^{-1}$) is given by

$$P_{disturb,j} = \frac{A_{b,j}}{f_j A_g} P_j \xi_j \quad (2.924)$$

where P_j (individual km^{-2}) is the population density for the j th tree PFT and ξ_j is the whole-plant mortality factor (Table 2.31).

2.23.2 Agricultural fires

The burned area of cropland ($\text{km}^2 \text{s}^{-1}$) is taken as A_b :

$$A_b = a_1 f_{se} f_t f_{crop} A_g \quad (2.925)$$

where a_1 (s^{-1}) is a constant; f_{se} represents the socioeconomic effect on fires; f_t determines the seasonality of agricultural fires; f_{crop} is the fractional coverage of cropland. $a_1 = 1.6 \times 10^{-4} \text{ hr}^{-1}$ is estimated using an inverse method, by matching 1997-2004 simulations to the analysis of [van der Werf et al. \(2010\)](#) that shows the 2001-2009 average contribution of cropland fires is 4.7% of the total global burned area.

The socioeconomic factor f_{se} is given as follows:

$$f_{se} = f_d f_e. \quad (2.926)$$

Here

$$f_d = 0.04 + 0.96 \times \exp\left[-\pi\left(\frac{D_p}{350}\right)^{0.5}\right] \quad (2.927)$$

and

$$f_e = 0.01 + 0.99 \times \exp\left(-\pi \frac{\text{GDP}}{10}\right) \quad (2.928)$$

are the effects of population density and GDP on burned area, derived in a similar way to equation (2.927) and (2.928). f_t is set to 1 at the first time step during the climatological peak month for agricultural fires ([van der Werf et al. 2010](#)); f_t is set to 0 otherwise. Peak month in this dataset correlates with the month after harvesting or the month before planting. In CLM we use this dataset the same way whether the CROP option is active or not, without regard to the CROP option's simulated planting and harvesting dates.

In the post-fire region, fire impact is parameterized similar to section 2.23.1 but with combustion completeness factors and tissue mortality factors for crop PFTs (Table 2.31).

2.23.3 Deforestation fires

CLM focuses on deforestation fires in tropical closed forests. Tropical closed forests are defined as grid cells with tropical tree (BET and BDT tropical) coverage > 60% according to the FAO classification. Deforestation fires are defined as fires caused by deforestation, including escaped deforestation fires, termed degradation fires. Deforestation and degradation fires are assumed to occur outside of cropland areas in these grid cells. Burned area is controlled by the deforestation rate and climate:

$$A_b = b f_{lu} f_{cli,d} f_b A_g \quad (2.929)$$

where b (s^{-1}) is a global constant; f_{lu} (fraction) represents the effect of decreasing fractional coverage of tree PFTs derived from land use data; $f_{cli,d}$ (fraction) represents the effect of climate conditions on the burned area.

Constants b and f_{lu} are calibrated based on observations and reanalysis datasets in the Amazon rainforest (tropical closed forests within $15.5^\circ S - 10.5^\circ N$, $30.5^\circ W - 91^\circ W$). $b = 0.033 d^{-1}$ and f_{lu} is defined as

$$f_{lu} = \max(0.0005, 0.19D - 0.001) \quad (2.930)$$

where D (yr^{-1}) is the annual loss of tree cover based on CLM land use and land cover change data.

The effect of climate on deforestation fires is parameterized as:

$$\begin{aligned} f_{cli,d} = & \max \left[0, \min \left(1, \frac{b_2 - P_{60d}}{b_2} \right) \right]^{0.5} \times \\ & \max \left[0, \min \left(1, \frac{b_3 - P_{10d}}{b_3} \right) \right]^{0.5} \times \\ & \max \left[0, \min \left(1, \frac{0.25 - P}{0.25} \right) \right] \end{aligned} \quad (2.931)$$

where P ($mm\ d^{-1}$) is instantaneous precipitation, while P_{60d} ($mm\ d^{-1}$) and P_{10d} ($mm\ d^{-1}$) are 60-day and 10-day running means of precipitation, respectively; b_2 ($mm\ d^{-1}$) and b_3 ($mm\ d^{-1}$) are the grid-cell dependent thresholds of P_{60d} and P_{10d} ; $0.25\ mm\ d^{-1}$ is the maximum precipitation rate for drizzle. [Le Page et al. \(2010\)](#) analyzed the relationship between large-scale deforestation fire counts and precipitation during 2003–2006 in southern Amazonia where tropical evergreen trees (BET Tropical) are dominant. Figure 2 in [Le Page et al. \(2010\)](#) showed that fires generally occurred if both P_{60d} and P_{10d} were less than about $4.0\ mm\ d^{-1}$, and fires occurred more frequently in a drier environment. Based on the 30-yr (1985 to 2004) precipitation data in [Qian et al. \(2006\)](#). The climatological precipitation of dry months ($P < 4.0\ mm\ d^{-1}$) in a year over tropical deciduous tree (BDT Tropical) dominated regions is 46% of that over BET Tropical dominated regions, so we set the PFT-dependent thresholds of P_{60d} and P_{10d} as $4.0\ mm\ d^{-1}$ for BET Tropical and $1.8\ mm\ d^{-1}$ ($= 4.0\ mm\ d^{-1} \times 46\%$) for BDT Tropical, and b_2 and b_3 are the average of thresholds of BET Tropical and BDT Tropical weighted by their coverage.

The post-fire area due to deforestation is not limited to land-type conversion regions. In the tree-reduced region, the maximum fire carbon emissions are assumed to be 80% of the total conversion flux. According to the fraction of conversion flux for tropical trees in the tree-reduced region (60%) assigned by CLM4-CN, to reach the maximum fire carbon emissions in a conversion region requires burning this region about twice when we set PFT-dependent combustion completeness factors to about 0.3 for stem [the mean of 0.2–0.4 used in [van der Werf et al. \(2010\)](#). Therefore, when the burned area calculated from equation (2.931) is no more than twice the tree-reduced area, we assume no escaped fires outside the land-type conversion region, and the fire-related fraction of the total conversion flux is estimated as $\frac{A_b/A_g}{2D}$. Otherwise, 80% of the total conversion flux is assumed to be fire carbon emissions, and the biomass combustion and vegetation mortality outside the tree-reduced regions with an area fraction of $\frac{A_b}{A_g} - 2D$ are set as in section 2.23.1.

2.23.4 Peat fires

The burned area due to peat fires is given as A_b :

$$A_b = c f_{cli,p} f_{peat} (1 - f_{sat}) A_g \quad (2.932)$$

where c (s^{-1}) is a constant; $f_{cli,p}$ represents the effect of climate on the burned area; f_{peat} is the fractional coverage of peatland in the grid cell; and f_{sat} is the fraction of the grid cell with a water table at the surface or higher. $c = 0.17 \times 10^{-3} \text{ hr}^{-1}$ for tropical peat fires and $c = 0.9 \times 10^{-5} \text{ hr}^{-1}$ for boreal peat fires are derived using an inverse method, by matching simulations to earlier studies: about 2.4 Mha peatland was burned over Indonesia in 1997 ([Page et al. 2002](#)) and the average burned area of peat fires in Western Canada was 0.2 Mha yr^{-1} for 1980-1999 ([Turetsky et al. 2004](#)).

For tropical peat fires, $f_{cli,p}$ is set as a function of long-term precipitation P_{60d} :

$$f_{cli,p} = \max \left[0, \min \left(1, \frac{4 - P_{60d}}{4} \right) \right]^2. \quad (2.933)$$

For boreal peat fires, $f_{cli,p}$ is set to

$$f_{cli,p} = \exp(-\pi \frac{\theta_{17cm}}{0.3}) \cdot \max[0, \min(1, \frac{T_{17cm} - T_f}{10})] \quad (2.934)$$

where θ_{17cm} is the wetness of the top 17 cm of soil.

Peat fires lead to peat burning and the combustion and mortality of vegetation over peatlands. For tropical peat fires, based on [Page et al. \(2002\)](#), about 6% of the peat carbon loss from stored carbon is caused by 33.9% of the peatland burned. Carbon emissions due to peat burning ($\text{g C m}^{-2} \text{ s}^{-1}$) are therefore set as the product of 6%/33.9%, burned area fraction of peat fire (s^{-1}), and soil organic carbon (g C m^{-2}). For boreal peat fires, the carbon emissions due to peat burning are set as 2.2 kg C m^{-2} peat fire area ([Turetsky et al. 2002](#)). Biomass combustion and vegetation mortality in post-fire peatlands are set the same as section 2.23.1 for non-crop PFTs and as section 2.23.2 for crops PFTs.

2.23.5 Fire trace gas and aerosol emissions

CESM2 is the first Earth system model that can model the full coupling among fire, fire emissions, land, and atmosphere. CLM5, as the land component of CESM2, calculates the surface trace gas and aerosol emissions due to fire and fire emission heights, as the inputs of atmospheric chemistry model and aerosol model.

Emissions for trace gas and aerosol species x and the j-th PFT, $E_{x,j}$ (g species s^{-1}), are given by

$$E_{x,j} = EF_{x,j} \frac{\phi_j}{[C]}. \quad (2.935)$$

Here, $EF_{x,j}$ ($\text{g species (g dm)}^{-1}$) is PFT-dependent emission factor scaled from biome-level values (Li et al., in prep, also used for FireMIP fire emissions data) by Dr. Val Martin and Dr. Li. $[C] = 0.5 \text{ (g C (g dm)}^{-1}$) is a conversion factor from dry matter to carbon.

Emission height is PFT-dependent: 4.3 km for needleleaf tree PFTs, 3 km for other boreal and temperate tree PFTs, 2.5 km for tropical tree PFTs, 2 km for shrub PFTs, and 1 km for grass and crop PFTs. These values are compiled from earlier studies by Dr. Val Martin.

Table 2.31: PFT-specific combustion completeness and fire mortality factors.

PFT	CC_{leaf}	CC_{stem}	CC_{root}	CC_{ts}	M_{leaf}	$M_{livestem,1}$	$M_{deadstem}$	M_{root}	M_{ts}	$M_{livestem,2}$	ξ_j
NET Temperate	0.80	0.30	0.00	0.50	0.80	0.15	0.15	0.15	0.50	0.35	0.15
NET Boreal	0.80	0.30	0.00	0.50	0.80	0.15	0.15	0.15	0.50	0.35	0.15
NDT Boreal	0.80	0.30	0.00	0.50	0.80	0.15	0.15	0.15	0.50	0.35	0.15
BET Tropical	0.80	0.27	0.00	0.45	0.80	0.13	0.13	0.13	0.45	0.32	0.13
BET Temperate	0.80	0.27	0.00	0.45	0.80	0.13	0.13	0.13	0.45	0.32	0.13
BDT Tropical	0.80	0.27	0.00	0.45	0.80	0.10	0.10	0.10	0.35	0.25	0.10
BDT Temperate	0.80	0.27	0.00	0.45	0.80	0.10	0.10	0.10	0.35	0.25	0.10
BDT Boreal	0.80	0.27	0.00	0.45	0.80	0.13	0.13	0.13	0.45	0.32	0.13
BES Temperate	0.80	0.35	0.00	0.55	0.80	0.17	0.17	0.17	0.55	0.38	0.17
BDS Temperate	0.80	0.35	0.00	0.55	0.80	0.17	0.17	0.17	0.55	0.38	0.17
BDS Boreal	0.80	0.35	0.00	0.55	0.80	0.17	0.17	0.17	0.55	0.38	0.17
C ₃ Grass Arctic	0.80	0.80	0.00	0.80	0.80	0.20	0.20	0.20	0.80	0.60	0.20
C ₃ Grass	0.80	0.80	0.00	0.80	0.80	0.20	0.20	0.20	0.80	0.60	0.20
C ₄ Grass	0.80	0.80	0.00	0.80	0.80	0.20	0.20	0.20	0.80	0.60	0.20
Crop	0.80	0.80	0.00	0.80	0.80	0.20	0.20	0.20	0.80	0.60	0.20

Leaves (CC_{leaf}), stems (CC_{stem}), roots (CC_{root}), and transfer and storage carbon (CC_{ts}); mortality factors for leaves (M_{leaf}), live stems ($M_{livestem,1}$), dead stems ($M_{deadstem}$), roots (M_{root}), and transfer and storage carbon (M_{ts}) related to the carbon transfers from these pools to litter pool; mortality factors for live stems ($M_{livestem,2}$) related to the carbon transfer from live stems to dead stems; whole-plant mortality factor (ξ_j).

2.24 Methane Model

The representation of processes in the methane biogeochemical model integrated in CLM [CLM4Me; (Riley et al. 2011a)] is based on several previously published models (Cao et al. 1996; Petrescu et al. 2010; Tianet al. 2010; Walter et al. 2001; Wania et al. 2010; Zhang et al. 2002; Zhuang et al. 2004). Although the model has similarities with these precursor models, a number of new process representations and parameterization have been integrated into CLM.

Mechanistically modeling net surface CH₄ emissions requires representing a complex and interacting series of processes. We first (section 2.24.1) describe the overall model structure and flow of information in the CH₄ model, then describe the methods used to represent: CH₄ mass balance; CH₄ production; ebullition; aerenchyma transport; CH₄ oxidation; reactive transport solution, including boundary conditions, numerical solution, water table interface, etc.; seasonal inundation effects; and impact of seasonal inundation on CH₄ production.

2.24.1 Methane Model Structure and Flow

The driver routine for the methane biogeochemistry calculations (ch4, in ch4Mod.F) controls the initialization of boundary conditions, inundation, and impact of redox conditions; calls to routines to calculate CH₄ production, oxidation, transport through aerenchyma, ebullition, and the overall mass balance (for unsaturated and saturated soils and, if

desired, lakes); resolves changes to CH₄ calculations associated with a changing inundated fraction; performs a mass balance check; and calculates the average gridcell CH₄ production, oxidation, and exchanges with the atmosphere.

2.24.2 Governing Mass-Balance Relationship

The model (Figure 2.19) accounts for CH₄ production in the anaerobic fraction of soil (P , mol m⁻³ s⁻¹), ebullition (E , mol m⁻³ s⁻¹), aerenchyma transport (A , mol m⁻³ s⁻¹), aqueous and gaseous diffusion (F_D , mol m⁻² s⁻¹), and oxidation (O , mol m⁻³ s⁻¹) via a transient reaction diffusion equation:

$$\frac{\partial(RC)}{\partial t} = \frac{\partial F_D}{\partial z} + P(z, t) - E(z, t) - A(z, t) - O(z, t) \quad (2.936)$$

Here z (m) represents the vertical dimension, t (s) is time, and R accounts for gas in both the aqueous and gaseous phases: $R = \epsilon_a + K_H \epsilon_w$, with ϵ_a , ϵ_w , and K_H (-) the air-filled porosity, water-filled porosity, and partitioning coefficient for the species of interest, respectively, and C represents CH₄ or O₂ concentration with respect to water volume (mol m⁻³).

An analogous version of equation is concurrently solved for O₂, but with the following differences relative to CH₄: $P = E = 0$ (i.e., no production or ebullition), and the oxidation sink includes the O₂ demanded by methanotrophs, heterotroph decomposers, nitrifiers, and autotrophic root respiration.

As currently implemented, each gridcell contains an inundated and a non-inundated fraction. Therefore, equation is solved four times for each gridcell and time step: in the inundated and non-inundated fractions, and for CH₄ and O₂. If desired, the CH₄ and O₂ mass balance equation is solved again for lakes (Chapter 9). For non-inundated areas, the water table interface is defined at the deepest transition from greater than 95% saturated to less than 95% saturated that occurs above frozen soil layers. The inundated fraction is allowed to change at each time step, and the total soil CH₄ quantity is conserved by evolving CH₄ to the atmosphere when the inundated fraction decreases, and averaging a portion of the non-inundated concentration into the inundated concentration when the inundated fraction increases.

2.24.3 CH₄ Production

Because CLM does not currently specifically represent wetland plant functional types or soil biogeochemical processes, we used gridcell-averaged decomposition rates as proxies. Thus, the upland (default) heterotrophic respiration is used to estimate the wetland decomposition rate after first dividing off the O₂ limitation. The O₂ consumption associated with anaerobic decomposition is then set to the unlimited version so that it will be reduced appropriately during O₂ competition. CH₄ production at each soil level in the anaerobic portion (i.e., below the water table) of the column is related to the gridcell estimate of heterotrophic respiration from soil and litter (R_H ; mol C m⁻² s⁻¹) corrected for its soil temperature (T_s) dependence, soil temperature through a A_{10} factor (f_T), pH (f_{pH}), redox potential (f_{pE}), and a factor accounting for the seasonal inundation fraction (S , described below):

$$P = R_H f_{CH_4} f_T f_{pH} f_{pE} S. \quad (2.937)$$

Here, f_{CH_4} is the baseline ratio between CO₂ and CH₄ production (all parameters values are given in Table 2.32). Currently, f_{CH_4} is modified to account for our assumptions that methanogens may have a higher Q₁₀ than aerobic decomposers; are not N limited; and do not have a low-moisture limitation.

When the single BGC soil level is used in CLM (Chapter 2.21), the temperature factor, f_T , is set to 0 for temperatures equal to or below freezing, even though CLM allows heterotrophic respiration below freezing. However, if the vertically resolved BGC soil column is used, CH₄ production continues below freezing because liquid water stress limits decomposition. The base temperature for the Q₁₀ factor, T_B , is 22° C and effectively modified the base f_{CH_4} value.

For the single-layer BGC version, R_H is distributed among soil levels by assuming that 50% is associated with the roots (using the CLM PFT-specific rooting distribution) and the rest is evenly divided among the top 0.28 m of soil (to be consistent with CLM's soil decomposition algorithm). For the vertically resolved BGC version, the prognosed distribution of R_H is used to estimate CH₄ production.

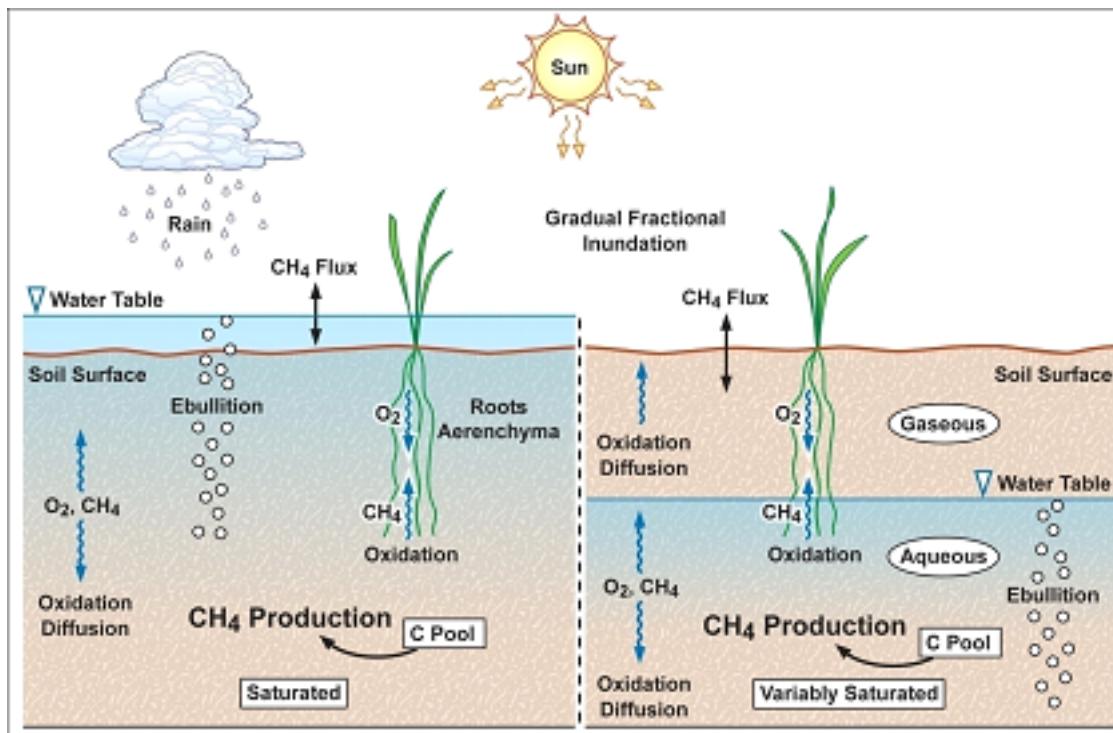


Figure 2.19: Schematic representation of biological and physical processes integrated in CLM that affect the net CH_4 surface flux (Riley et al. 2011a). (left) Fully inundated portion of a CLM gridcell and (right) variably saturated portion of a gridcell.

The factor f_{pH} is nominally set to 1, although a static spatial map of pH can be used to determine this factor (*Dunfield et al. 1993*) by applying:

$$f_{pH} = 10^{-0.2235pH^2 + 2.7727pH - 8.6}. \quad (2.938)$$

The f_{pE} factor assumes that alternative electron acceptors are reduced with an e-folding time of 30 days after inundation. The default version of the model applies this factor to horizontal changes in inundated area but not to vertical changes in the water table depth in the upland fraction of the gridcell. We consider both f_{pH} and f_{pE} to be poorly constrained in the model and identify these controllers as important areas for model improvement.

As a non-default option to account for CH₄ production in anoxic microsites above the water table, we apply the Arah and Stephen (1998) estimate of anaerobic fraction:

$$\varphi = \frac{1}{1 + \eta C_{O_2}}. \quad (2.939)$$

Here, φ is the factor by which production is inhibited above the water table (compared to production as calculated in equation), C_{O_2} (mol m⁻³) is the bulk soil oxygen concentration, and $\eta = 400$ mol m⁻³.

The O₂ required to facilitate the vertically resolved heterotrophic decomposition and root respiration is estimated assuming 1 mol O₂ is required per mol CO₂ produced. The model also calculates the O₂ required during nitrification, and the total O₂ demand is used in the O₂ mass balance solution.

Table 2.32: Parameter descriptions and sensitivity analysis ranges applied in the methane model

Mechanism	Parameter	Baseline Value	Range for Sensitivity Analysis	Units	Description
Production	Q_{10}	2	1.5 – 4	•	CH ₄ production Q_{10}
	f_{pH}	1	On, off	•	Impact of pH on CH ₄ production
	f_{pE}	1	On, off	•	Impact of redox potential on CH ₄ production
	S	Varies	NA	•	Seasonal inundation factor
	β	0.2	NA	•	Effect of anoxia on decomposition rate (used to calculate S only)
	f_{CH_4}	0.2	NA	•	Ratio between CH ₄ and CO ₂ production below the water table
Ebullition	$C_{e,max}$	0.15	NA	mol m ⁻³	CH ₄ concentration to start ebullition
	$C_{e,min}$	0.15	NA	•	CH ₄ concentration to end ebullition
Diffusion	f_{D_0}	1	1, 10	m ² s ⁻¹	Diffusion coefficient multiplier (Table 24.2)
Aerenchyma	p	0.3	NA	•	Grass aerenchyma porosity
	R	2.9×10^{-3} m	NA	m	Aerenchyma radius
	r_L	3	NA	•	Root length to depth ratio
	F_a	1	0.5 – 1.5	•	Aerenchyma conductance multiplier
Oxidation	K_{CH_4}	5×10^{-3}	5×10^{-4} - 5×10^{-2}	mol m ⁻³	CH ₄ half-saturation oxidation coefficient (wetlands)
	K_{O_2}	2×10^{-2}	2×10^{-3} - 2×10^{-1}	mol m ⁻³	O ₂ half-saturation oxidation coefficient
2.24. Methane Model	$K_{O_2,max}$	1.25×10^{-5}	1.25×10^{-6} - 1.25×10^{-4}	mol m ⁻³ s ⁻¹	Maximum oxidation rate (wetlands) 283

2.24.4 Ebullition

Briefly, the simulated aqueous CH₄ concentration in each soil level is used to estimate the expected equilibrium gaseous partial pressure (C_e), as a function of temperature and depth below the water table, by first estimating the Henry's law partitioning coefficient (k_h^C) by the method described in [Wania et al. \(2010\)](#):

$$\log\left(\frac{1}{k_H}\right) = \log k_H^s - \frac{1}{C_H} \left(\frac{1}{T} - \frac{1}{T^s} \right) \quad (2.940)$$

$$k_h^C = Tk_H R_g \quad (2.941)$$

$$C_e = \frac{C_w R_g T}{\theta_s k_H^C p} \quad (2.942)$$

where C_H is a constant, R_g is the universal gas constant, k_H^s is Henry's law partitioning coefficient at standard temperature (T^s), C_w is local aqueous CH₄ concentration, and p is pressure.

The local pressure is calculated as the sum of the ambient pressure, water pressure down to the local depth, and pressure from surface ponding (if applicable). When the CH₄ partial pressure exceeds 15% of the local pressure (Baird et al. 2004; Strack et al. 2006; Wania et al. 2010), bubbling occurs to remove CH₄ to below this value, modified by the fraction of CH₄ in the bubbles [taken as 57%; ([Kellner et al. 2006](#); [Wania et al. 2010](#))]. Bubbles are immediately added to the surface flux for saturated columns and are placed immediately above the water table interface in unsaturated columns.

2.24.5 Aerenchyma Transport

Aerenchyma transport is modeled in CLM as gaseous diffusion driven by a concentration gradient between the specific soil layer and the atmosphere and, if specified, by vertical advection with the transpiration stream. There is evidence that pressure driven flow can also occur, but we did not include that mechanism in the current model.

The diffusive transport through aerenchyma (A , mol m⁻² s⁻¹) from each soil layer is represented in the model as:

$$A = \frac{C(z) - C_a}{r_L z / D + r_a} p T \rho_r, \quad (2.943)$$

where D is the free-air gas diffusion coefficient (m:²s⁻¹); $C(z)$ (mol m⁻³) is the gaseous concentration at depth z (m); r_L is the ratio of root length to depth; p is the porosity (-); T is specific aerenchyma area (m:²m⁻²); r_a is the aerodynamic resistance between the surface and the atmospheric reference height (s m:²); and ρ_r is the rooting density as a function of depth (-). The gaseous concentration is calculated with Henry's law as described in equation .

Based on the ranges reported in [Colmer \(2003\)](#), we have chosen baseline aerenchyma porosity values of 0.3 for grass and crop PFTs and 0.1 for tree and shrub PFTs:

$$T = \frac{4f_N N_a}{0.22} \pi R^2. \quad (2.944)$$

Here N_a is annual net primary production (NPP, mol m⁻² s⁻¹); R is the aerenchyma radius (2.9×10^{-3} m); f_N is the belowground fraction of annual NPP; and the 0.22 factor represents the amount of C per tiller. O₂ can also diffuse in from the atmosphere to the soil layer via the reverse of the same pathway, with the same representation as Equation but with the gas diffusivity of oxygen.

CLM also simulates the direct emission of CH₄ from leaves to the atmosphere via transpiration of dissolved methane. We calculate this flux (F_{CH_4-T} ; mol m⁻² s⁻¹) using the simulated soil water methane concentration ($C_{CH_4,j}$ (mol m⁻³)) in each soil layer j and the CLM predicted transpiration (F_T) for each PFT, assuming that no methane was oxidized inside the plant tissue:

$$F_{CH_4-T} = \sum_j \rho_{r,j} F_T C_{CH_4,j}. \quad (2.945)$$

2.24.6 CH₄ Oxidation

CLM represents CH₄ oxidation with double Michaelis-Menten kinetics ([Arah and Stephen 1998; Segers 1998](#)), dependent on both the gaseous CH₄ and O₂ concentrations:

$$R_{oxic} = R_{o,max} \left[\frac{C_{CH_4}}{K_{CH_4} + C_{CH_4}} \right] \left[\frac{C_{O_2}}{K_{O_2} + C_{O_2}} \right] Q_{10} F_\theta \quad (2.946)$$

where K_{CH_4} and K_{O_2} are the half saturation coefficients (mol m⁻³) with respect to CH₄ and O₂ concentrations, respectively; $R_{o,max}$ is the maximum oxidation rate (mol m⁻³ s⁻¹); and Q_{10} specifies the temperature dependence of the reaction with a base temperature set to 12 ° C. The soil moisture limitation factor F_θ is applied above the water table to represent water stress for methanotrophs. Based on the data in [Schnell and King \(1996\)](#), we take $F_\theta = e^{-P/P_c}$, where P is the soil moisture potential and $P_c = -2.4 \times 10^5$ mm.

2.24.7 Reactive Transport Solution

The solution to equation is solved in several sequential steps: resolve competition for CH₄ and O₂ (section 2.24.7); add the ebullition flux into the layer directly above the water table or into the atmosphere; calculate the overall CH₄ or O₂ source term based on production, aerenchyma transport, ebullition, and oxidation; establish boundary conditions, including surface conductance to account for snow, ponding, and turbulent conductances and bottom flux condition (section 2.24.7); calculate diffusivity (section 2.24.7); and solve the resulting mass balance using a tridiagonal solver (section 2.24.7).

Competition for CH₄ and O₂

For each time step, the unlimited CH₄ and O₂ demands in each model depth interval are computed. If the total demand over a time step for one of the species exceeds the amount available in a particular control volume, the demand from each process associated with the sink is scaled by the fraction required to ensure non-negative concentrations. Since the methanotrophs are limited by both CH₄ and O₂, the stricter limitation is applied to methanotroph oxidation, and then the limitations are scaled back for the other processes. The competition is designed so that the sinks must not exceed the available concentration over the time step, and if any limitation exists, the sinks must sum to this value. Because the sinks are calculated explicitly while the transport is semi-implicit, negative concentrations can occur after the tridiagonal solution. When this condition occurs for O₂, the concentrations are reset to zero; if it occurs for CH₄, the surface flux is adjusted and the concentration is set to zero if the adjustment is not too large.

CH₄ and O₂ Source Terms

The overall CH₄ net source term consists of production, oxidation at the base of aerenchyma, transport through aerenchyma, methanotrophic oxidation, and ebullition (either to the control volume above the water table if unsaturated or directly to the atmosphere if saturated). For O₂ below the top control volume, the net source term consists of O₂ losses from methanotropy, SOM decomposition, and autotrophic respiration, and an O₂ source through aerenchyma.

Aqueous and Gaseous Diffusion

For gaseous diffusion, we adopted the temperature dependence of molecular free-air diffusion coefficients (D_0 (m²:sup:2 s⁻¹)) as described by [Lerman \(1979\)](#) and applied by [Wania et al. \(2010\)](#) (Table 2.33).

Table 2.33: Temperature dependence of aqueous and gaseous diffusion coefficients for CH₄ and O₂

D_0 (m ² s ⁻¹)	CH ₄	O ₂
Aqueous	$0.9798 + 0.02986T + 0.0004381T^2$	$1.172 + 0.03443T + 0.0005048T^2$
Gaseous	$0.1875 + 0.0013T$	$0.1759 + 0.0011T$

Gaseous diffusivity in soils also depends on the molecular diffusivity, soil structure, porosity, and organic matter content. [Moldrup et al. \(2003\)](#), using observations across a range of unsaturated mineral soils, showed that the relationship between effective diffusivity (D_e (m² s⁻¹)) and soil properties can be represented as:

$$D_e = D_0 \theta_a^2 \left(\frac{\theta_a}{\theta_s} \right)^{3/b}, \quad (2.947)$$

where θ_a and θ_s are the air-filled and total (saturated water-filled) porosities (-), respectively, and b is the slope of the water retention curve (-). However, [Iiyama and Hasegawa \(2005\)](#) have shown that the original Millington-Quirk ([Millington and Quirk 1961](#)) relationship matched measurements more closely in unsaturated peat soils:

$$D_e = D_0 \frac{\theta_a^{10/3}}{\theta_s^2} \quad (2.948)$$

In CLM, we applied equation for soils with zero organic matter content and equation for soils with more than 130 kg m⁻³ organic matter content. A linear interpolation between these two limits is applied for soils with SOM content below 130 kg m⁻³. For aqueous diffusion in the saturated part of the soil column, we applied ([Moldrup et al. \(2003\)](#)):

$$D_e = D_0 \theta_s^2. \quad (2.949)$$

To simplify the solution, we assumed that gaseous diffusion dominates above the water table interface and aqueous diffusion below the water table interface. Descriptions, baseline values, and dimensions for parameters specific to the CH₄ model are given in [Table 2.32](#). For freezing or frozen soils below the water table, diffusion is limited to the remaining liquid (CLM allows for some freezing point depression), and the diffusion coefficients are scaled by the volume-fraction of liquid. For unsaturated soils, Henry's law equilibrium is assumed at the interface with the water table.

Boundary Conditions

We assume the CH₄ and O₂ surface fluxes can be calculated from an effective conductance and a gaseous concentration gradient between the atmospheric concentration and either the gaseous concentration in the first soil layer (unsaturated soils) or in equilibrium with the water (saturated soil w ($C_1^n - C_a$) and w ($C_1^{n+1} - C_a$) for the fully explicit and fully implicit cases, respectively (however, see [Tang and Riley \(2013\)](#) for a more complete representation of this process). Here, w is the surface boundary layer conductance as calculated in the existing CLM surface latent heat calculations.

If the top layer is not fully saturated, the $\frac{D_{m1}}{\Delta x_{m1}}$ term is replaced with a series combination: $\left[\frac{1}{w} + \frac{\Delta x_1}{D_1} \right]^{-1}$, and if the top layer is saturated, this term is replaced with $\left[\frac{K_H}{w} + \frac{\frac{1}{2} \Delta x_1}{D_1} \right]^{-1}$, where K_H is the Henry's law equilibrium constant.

When snow is present, a resistance is added to account for diffusion through the snow based on the Millington-Quirk expression (2.948) and CLM's prediction of the liquid water, ice, and air fractions of each snow layer. When the soil is ponded, the diffusivity is assumed to be that of methane in pure water, and the resistance as the ratio of the ponding depth to diffusivity. The overall conductance is taken as the series combination of surface, snow, and ponding resistances. We assume a zero flux gradient at the bottom of the soil column.

Crank-Nicholson Solution

Equation is solved using a Crank-Nicholson solution ([Press et al. 1992](#)), which combines fully explicit and implicit representations of the mass balance. The fully explicit decomposition of equation can be written as

$$\frac{R_j^{n+1} C_j^{n+1} - R_j^n C_j^n}{\Delta t} = \frac{1}{\Delta x_j} \left[\frac{D_{p1}^n}{\Delta x_{p1}} (C_{j+1}^n - C_j^n) - \frac{D_{m1}^n}{\Delta x_{m1}} (C_j^n - C_{j-1}^n) \right] + S_j^n, \quad (2.950)$$

where j refers to the cell in the vertically discretized soil column (increasing downward), n refers to the current time step (s), $p1$ is $j+1/2$, $m1$ is $j-1/2$, and S_j^n is the net source at time step n and position j , i.e.,

$S_j^n = P(j, n) - E(j, n) - A(j, n) - O(j, n)$. The diffusivity coefficients are calculated as harmonic means of values from the adjacent cells. Equation is solved for gaseous and aqueous concentrations above and below the water table, respectively. The R term ensure the total mass balance in both phases is properly accounted for. An analogous relationship can be generated for the fully implicit case by replacing n by $n+1$ on the C and S terms of equation . Using an average of the fully implicit and fully explicit relationships gives:

$$\begin{aligned} -\frac{1}{2\Delta x_j} \frac{D_{m1}}{\Delta x_{m1}} C_{j-1}^{n+1} + \left[\frac{R_j^{n+1}}{\Delta t} + \frac{1}{2\Delta x_j} \left(\frac{D_{p1}}{\Delta x_{p1}} + \frac{D_{m1}}{\Delta x_{m1}} \right) \right] C_j^{n+1} - \frac{1}{2\Delta x_j} \frac{D_{p1}}{\Delta x_{p1}} C_{j+1}^{n+1} = \\ \frac{R_j^n}{\Delta t} + \frac{1}{2\Delta x_j} \left[\frac{D_{p1}}{\Delta x_{p1}} (C_{j+1}^n - C_j^n) - \frac{D_{m1}}{\Delta x_{m1}} (C_j^n - C_{j-1}^n) \right] + \frac{1}{2} [S_j^n + S_j^{n+1}] \end{aligned} \quad (2.951)$$

Equation is solved with a standard tridiagonal solver, i.e.:

$$aC_{j-1}^{n+1} + bC_j^{n+1} + cC_{j+1}^{n+1} = r, \quad (2.952)$$

with coefficients specified in equation .

Two methane balance checks are performed at each timestep to insure that the diffusion solution and the time-varying aggregation over inundated and non-inundated areas strictly conserves methane molecules (except for production minus consumption) and carbon atoms.

Interface between water table and unsaturated zone

We assume Henry's Law equilibrium at the interface between the saturated and unsaturated zone and constant flux from the soil element below the interface to the center of the soil element above the interface. In this case, the coefficients are the same as described above, except for the soil element above the interface:

$$\begin{aligned} \frac{D_{p1}}{\Delta x_{p1}} &= \left[K_H \frac{\Delta x_j}{2D_j} + \frac{\Delta x_{j+1}}{2D_{j+1}} \right]^{-1} \\ b &= \left[\frac{R_j^{n+1}}{\Delta t} + \frac{1}{2\Delta x_j} \left(K_H \frac{D_{p1}}{\Delta x_{p1}} + \frac{D_{m1}}{\Delta x_{m1}} \right) \right] \\ r &= \frac{R_j^n}{\Delta t} C_j^n + \frac{1}{2\Delta x_j} \left[\frac{D_{p1}}{\Delta x_{p1}} (C_{j+1}^n - K_H C_j^n) - \frac{D_{m1}}{\Delta x_{m1}} (C_j^n - C_{j-1}^n) \right] + \frac{1}{2} [S_j^n + S_j^{n+1}] \end{aligned} \quad (2.953)$$

and the soil element below the interface:

$$\begin{aligned} \frac{D_{m1}}{\Delta x_{m1}} &= \left[K_H \frac{\Delta x_{j-1}}{2D_{j-1}} + \frac{\Delta x_j}{2D_j} \right]^{-1} \\ a &= -K_H \frac{1}{2\Delta x_j} \frac{D_{m1}}{\Delta x_{m1}} \\ r &= \frac{R_j^n}{\Delta t} + C_j^n + \frac{1}{2\Delta x_j} \left[\frac{D_{p1}}{\Delta x_{p1}} (C_{j+1}^n - C_j^n) - \frac{D_{m1}}{\Delta x_{m1}} (C_j^n - K_H C_{j-1}^n) \right] + \frac{1}{2} [S_j^n + S_j^{n+1}] \end{aligned} \quad (2.954)$$

2.24.8 Inundated Fraction Prediction

A simplified dynamic representation of spatial inundation based on recent work by [Prigent et al. \(2007\)](#) is used. Prigent et al. (2007) described a multi-satellite approach to estimate the global monthly inundated fraction (F_i) over an equal area grid ($0.25^\circ \times 0.25^\circ$ at the equator) from 1993 - 2000. They suggested that the IGBP estimate for inundation could be used as a measure of sensitivity of their detection approach at low inundation. We therefore used

the sum of their satellite-derived F_i and the constant IGBP estimate when it was less than 10% to perform a simple inversion for the inundated fraction for methane production (f_s). The method optimized two parameters (fws_{slope} and $fws_{intercept}$) for each grid cell in a simple model based on simulated total water storage (TWS):

$$f_s = fws_{slope}TWS + fws_{intercept}. \quad (2.955)$$

These parameters were evaluated at the 0.5° resolution, and aggregated for coarser simulations. Ongoing work in the hydrology submodel of CLM may alleviate the need for this crude simplification of inundated fraction in future model versions.

2.24.9 Seasonal Inundation

A simple scaling factor is used to mimic the impact of seasonal inundation on CH₄ production (see appendix B in *Riley et al. (2011a)* for a discussion of this simplified expression):

$$S = \frac{\beta(f - \bar{f}) + \bar{f}}{f}, S \leq 1. \quad (2.956)$$

Here, f is the instantaneous inundated fraction, \bar{f} is the annual average inundated fraction (evaluated for the previous calendar year) weighted by heterotrophic respiration, and β is the anoxia factor that relates the fully anoxic decomposition rate to the fully oxygen-unlimited decomposition rate, all other conditions being equal.

2.25 Crops and Irrigation

2.25.1 Summary of CLM5.0 updates relative to the CLM4.5

We describe here the complete crop and irrigation parameterizations that appear in CLM5.0. Corresponding information for CLM4.5 appeared in the CLM4.5 Technical Note (*Oleson et al. 2013*).

CLM5.0 includes the following new updates to the CROP option, where CROP refers to the interactive crop management model and is included as an option with the BGC configuration:

- New crop functional types
- All crop areas are actively managed
- Fertilization rates updated based on crop type and geographic region
- New Irrigation triggers
- Phenological triggers vary by latitude for some crop types
- Ability to simulate transient crop management
- Adjustments to allocation and phenological parameters
- Crops reaching their maximum LAI triggers the grain fill phase
- Grain C and N pools are included in a 1-year product pool
- C for annual crop seeding comes from the grain C pool
- Initial seed C for planting is increased from 1 to 3 g C/m²

These updates appear in detail in the sections below. Many also appear in *Levis et al. (2016)*.

2.25.2 The crop model

Introduction

Groups developing Earth System Models generally account for the human footprint on the landscape in simulations of historical and future climates. Traditionally we have represented this footprint with natural vegetation types and particularly grasses because they resemble many common crops. Most modeling efforts have not incorporated more explicit representations of land management such as crop type, planting, harvesting, tillage, fertilization, and irrigation, because global scale datasets of these factors have lagged behind vegetation mapping. As this begins to change, we increasingly find models that will simulate the biogeophysical and biogeochemical effects not only of natural but also human-managed land cover.

AgroIBIS is a state-of-the-art land surface model with options to simulate dynamic vegetation ([Kucharik et al. 2000](#)) and interactive crop management ([Kucharik and Brye 2003](#)). The interactive crop management parameterizations from AgroIBIS (March 2003 version) were coupled as a proof-of-concept to the Community Land Model version 3 [CLM3.0, [Oleson et al. \(2004\)](#)] (not published), then coupled to the CLM3.5 ([Levis et al. 2009](#)) and later released to the community with CLM4CN ([Levis et al. 2012](#)), and CLM4.5BGC. Additional updates after the release of CLM4.5 were available by request ([Levis et al. 2016](#)), and those are now incorporated into CLM5.

With interactive crop management and, therefore, a more accurate representation of agricultural landscapes, we hope to improve the CLM's simulated biogeophysics and biogeochemistry. These advances may improve fully coupled simulations with the Community Earth System Model (CESM), while helping human societies answer questions about changing food, energy, and water resources in response to climate, environmental, land use, and land management change (e.g., [Kucharik and Brye 2003](#); [Lobell et al. 2006](#)). As implemented here, the crop model uses the same physiology as the natural vegetation, though uses different crop-specific parameter values, phenology, and allocation, as well as fertilizer and irrigation management.

Crop plant functional types

To allow crops to coexist with natural vegetation in a grid cell, the vegetated land unit is separated into a naturally vegetated land unit and a managed crop land unit. Unlike the plant functional types (pfts) in the naturally vegetated land unit, the managed crop pfts in the managed crop land unit do not share soil columns and thus permit for differences in the land management between crops. Each crop type has a rainfed and an irrigated pft that are on independent soil columns. Crop grid cell coverage is assigned from satellite data (similar to all natural pfts), and the managed crop type proportions within the crop area is based on the dataset created by [Portmann et al. \(2010\)](#) for present day. New in CLM5, crop area is extrapolated through time using the dataset provided by Land Use Model Intercomparison Project (LUMIP), which is part of CMIP6 Land use timeseries ([Lawrence et al. 2016](#)). For more details about how crop distributions are determined, see Chapter 2.26.

CLM5 includes eight actively managed crop types (temperate soybean, tropical soybean, temperate corn, tropical corn, spring wheat, cotton, rice, and sugarcane) that are chosen based on the availability of corresponding algorithms in AgroIBIS and as developed by [Badger and Dirmeyer \(2015\)](#) and described by [Levis et al. \(2016\)](#). The representations of sugarcane, rice, cotton, tropical corn, and tropical soy are new in CLM5. Sugarcane and tropical corn are both C4 plants and are therefore represented using the temperate corn functional form. Tropical soybean uses the temperate soybean functional form, while rice and cotton use the wheat functional form. In tropical regions, parameter values were developed for the Amazon Basin, and planting date window is shifted by six months relative to the Northern Hemisphere.

In addition, CLM's default list of plant functional types (pfts) includes an irrigated and unirrigated unmanaged C3 crop ([Table 2.34](#)) treated as a second C3 grass. The unmanaged C3 crop is only used when the crop model is not active and has grid cell coverage assigned from satellite data, and the unmanaged C3 irrigated crop type is currently not used since irrigation requires the crop model to be active. The default list of pfts also includes twenty-three inactive crop pfts that do not yet have associated parameters required for active management. Each of the inactive crop types is simulated using the parameters of the spatially closest associated crop type that is most similar to the functional type (e.g., C3 or C4), which is required to maintain similar phenological parameters based on temperature thresholds.

Information detailing which parameters are used for each crop type is included in [Table 2.34](#). It should be noted that pft-level history output merges all crop types into the actively managed crop type, so analysis of crop-specific output will require use of the land surface dataset to remap the yields of each actively and inactively managed crop type. Otherwise, the actively managed crop type will include yields for that crop type and all inactively managed crop types that are using the same parameter set.

Table 2.34: Crop plant functional types (pfts) included in CLM5BGCCROP.

ITV	Plant function types (PFTs)	Management Class	Crop Parameters Used
15	c3 unmanaged rainfed crop	none	not applicable
16	c3 unmanaged irrigated crop	none	not applicable
17	rainfed temperate corn	active	rainfed temperate corn
18	irrigated temperate corn	active	irrigated temperate corn
19	rainfed spring wheat	active	rainfed spring wheat
20	irrigated spring wheat	active	irrigated spring wheat
21	rainfed winter wheat	inactive	rainfed spring wheat
22	irrigated winter wheat	inactive	irrigated spring wheat
23	rainfed temperate soybean	active	rainfed temperate soybean
24	irrigated temperate soybean	active	irrigated temperate soybean
25	rainfed barley	inactive	rainfed spring wheat
26	irrigated barley	inactive	irrigated spring wheat
27	rainfed winter barley	inactive	rainfed spring wheat
28	irrigated winter barley	inactive	irrigated spring wheat
29	rainfed rye	inactive	rainfed spring wheat
30	irrigated rye	inactive	irrigated spring wheat
31	rainfed winter rye	inactive	rainfed spring wheat
32	irrigated winter rye	inactive	irrigated spring wheat
33	rainfed cassava	inactive	rainfed rice
34	irrigated cassava	inactive	irrigated rice
35	rainfed citrus	inactive	rainfed spring wheat
36	irrigated citrus	inactive	irrigated spring wheat
37	rainfed cocoa	inactive	rainfed rice
38	irrigated cocoa	inactive	irrigated rice
39	rainfed coffee	inactive	rainfed rice
40	irrigated coffee	inactive	irrigated rice
41	rainfed cotton	active	rainfed cotton
42	irrigated cotton	active	irrigated cotton
43	rainfed datepalm	inactive	rainfed cotton
44	irrigated datepalm	inactive	irrigated cotton
45	rainfed foddergrass	inactive	rainfed spring wheat
46	irrigated foddergrass	inactive	irrigated spring wheat
47	rainfed grapes	inactive	rainfed spring wheat
48	irrigated grapes	inactive	irrigated spring wheat
49	rainfed groundnuts	inactive	rainfed rice
50	irrigated groundnuts	inactive	irrigated rice
51	rainfed millet	inactive	rainfed tropical corn
52	irrigated millet	inactive	irrigated tropical corn
53	rainfed oilpalm	inactive	rainfed rice
54	irrigated oilpalm	inactive	irrigated rice
55	rainfed potatoes	inactive	rainfed spring wheat
56	irrigated potatoes	inactive	irrigated spring wheat

Continued on next page

Table 2.34 – continued from previous page

ITV	Plant function types (PFTs)	Management Class	Crop Parameters Used
57	rainfed pulses	inactive	rainfed spring wheat
58	irrigated pulses	inactive	irrigated spring wheat
59	rainfed rapeseed	inactive	rainfed spring wheat
60	irrigated rapeseed	inactive	irrigated spring wheat
61	rainfed rice	active	rainfed rice
62	irrigated rice	active	irrigated rice
63	rainfed sorghum	inactive	rainfed tropical corn
64	irrigated sorghum	inactive	irrigated tropical corn
65	rainfed sugarbeet	inactive	rainfed spring wheat
66	irrigated sugarbeet	inactive	irrigated spring wheat
67	rainfed sugarcane	active	rainfed sugarcane
68	irrigated sugarcane	active	irrigated sugarcane
69	rainfed sunflower	inactive	rainfed spring wheat
70	irrigated sunflower	inactive	irrigated spring wheat
71	rainfed miscanthus	inactive	rainfed tropical corn
72	irrigated miscanthus	inactive	irrigated tropical corn
73	rainfed switchgrass	inactive	rainfed tropical corn
74	irrigated switchgrass	inactive	irrigated tropical corn
75	rainfed tropical corn	active	rainfed tropical corn
76	irrigated tropical corn	active	irrigated tropical corn
77	rainfed tropical soybean	active	rainfed tropical soybean
78	irrigated tropical soybean	active	irrigated tropical soybean

Phenology

CLM5-BGC includes evergreen, seasonally deciduous (responding to changes in day length), and stress deciduous (responding to changes in temperature and/or soil moisture) phenology algorithms (Chapter 2.20). CLM5-BGC-crop uses the AgroIBIS crop phenology algorithm, consisting of three distinct phases.

Phase 1 starts at planting and ends with leaf emergence, phase 2 continues from leaf emergence to the beginning of grain fill, and phase 3 starts from the beginning of grain fill and ends with physiological maturity and harvest.

Planting

All crops must meet the following requirements between the minimum planting date and the maximum planting date (for the northern hemisphere) in Table 2.35:

$$\begin{aligned} T_{10d} &> T_p \\ T_{10d}^{\min} &> T_p^{\min} \\ GDD_8 &\geq GDD_{\min} \end{aligned} \quad (2.957)$$

where T_{10d} is the 10-day running mean of T_{2m} , (the simulated 2-m air temperature during each model time step) and T_{10d}^{\min} is the 10-day running mean of T_{2m}^{\min} (the daily minimum of T_{2m}). T_p and T_p^{\min} are crop-specific coldest planting temperatures (Table 2.35), GDD_8 is the 20-year running mean growing degree-days (units are degree-days or $^{\circ}$ days) tracked from April through September (NH) above 8° C with maximum daily increments of 30° days (see equation (2.959)), and GDD_{\min} is the minimum growing degree day requirement (Table 2.35). GDD_8 does not change as quickly as T_{10d} and T_{10d}^{\min} , so it determines whether it is warm enough for the crop to be planted in a grid cell, while the 2-m air temperature variables determine the day when the crop may be planted if the GDD_8 threshold is met. If the requirements in equation (2.957) are not met by the maximum planting date, crops are still planted on the maximum planting date as long as $GDD_8 > 0$. In the southern hemisphere (SH) the NH requirements apply 6 months later.

At planting, each crop seed pool is assigned 3 gC m^{-2} from its grain product pool. The seed carbon is transferred to the leaves upon leaf emergence. An equivalent amount of seed leaf N is assigned given the pft's C to N ratio for leaves (CN_{leaf} in [Table 2.36](#); this differs from AgroIBIS, which uses a seed leaf area index instead of seed C). The model updates the average growing degree-days necessary for the crop to reach vegetative and physiological maturity, GDD_{mat} , according to the following AgroIBIS rules:

$$\begin{aligned} GDD_{mat}^{\text{corn,sugarcane}} &= 0.85GDD_8 && \text{and } 950 < GDD_{mat}^{\text{corn,sugarcane}} < 1850^\circ\text{days} \\ GDD_{mat}^{\text{spring wheat,cotton}} &= GDD_0 && \text{and } GDD_{mat}^{\text{spring wheat,cotton}} < 1700^\circ\text{days} \\ GDD_{mat}^{\text{temp.soy}} &= GDD_{10} && \text{and } GDD_{mat}^{\text{temp.soy}} < 1900^\circ\text{days} \\ GDD_{mat}^{\text{rice}} &= GDD_0 && \text{and } GDD_{mat}^{\text{rice}} < 2100^\circ\text{days} \\ GDD_{mat}^{\text{trop.soy}} &= GDD_{10} && \text{and } GDD_{mat}^{\text{trop.soy}} < 2100^\circ\text{days} \end{aligned} \quad (2.958)$$

where GDD_0 , GDD_8 , and GDD_{10} are the 20-year running mean growing degree-days tracked from April through September (NH) over 0°C , 8°C , and 10°C , respectively, with maximum daily increments of 26°days (for GDD_0) or 30°days (for GDD_8 and GDD_{10}). Equation (2.959) shows how we calculate GDD_0 , GDD_8 , and GDD_{10} for each model timestep:

$$\begin{aligned} GDD_0 &= GDD_0 + T_{2m} - T_f && \text{where } 0 \leq T_{2m} - T_f \leq 26^\circ\text{days} \\ GDD_8 &= GDD_8 + T_{2m} - T_f - 8 && \text{where } 0 \leq T_{2m} - T_f - 8 \leq 30^\circ\text{days} \\ GDD_{10} &= GDD_{10} + T_{2m} - T_f - 10 && \text{where } 0 \leq T_{2m} - T_f - 10 \leq 30^\circ\text{days} \end{aligned} \quad (2.959)$$

where, if $T_{2m} - T_f$ takes on values outside the above ranges within a day, then it equals the minimum or maximum value in the range for that day. T_f is the freezing temperature of water and equals 273.15 K , T_{2m} is the 2-m air temperature in units of K , and GDD is in units of $^\circ\text{days}$.

Leaf emergence

According to AgroIBIS, leaves may emerge when the growing degree-days of soil temperature to 0.05 m depth ($GDD_{T_{soi}}$), which is tracked since planting, reaches 1 to 5% of GDD_{mat} (see Phase 2 % GDD_{mat} in [Table 2.35](#)). The base temperature threshold values for $GDD_{T_{soi}}$ are listed in [Table 2.35](#) (the same base temperature threshold values are also used for $GDD_{T_{2m}}$ in section [2.25.2](#)), and leaf emergence (crop phenology phase 2) starts when this threshold is met. Leaf onset occurs in the first time step of phase 2, at which moment all seed C is transferred to leaf C. Subsequently, the leaf area index generally increases throughout phase 2 until it reaches a predetermined maximum value. Stem and root C also increase throughout phase 2 based on the carbon allocation algorithm in section [2.25.2](#).

Grain fill

The grain fill phase (phase 3) begins in one of two ways. The first potential trigger is based on temperature, similar to phase 2. A variable tracked since planting, similar to $GDD_{T_{soi}}$ but for 2-m air temperature, $GDD_{T_{2m}}$, must reach a heat unit threshold, h , of 40 to 65% of GDD_{mat} (see Phase 3 % GDD_{mat} in [Table 2.35](#)). For crops with the C4 photosynthetic pathway (temperate and tropical corn, sugarcane), the GDD_{mat} is based on an empirical function and ranges between 950 and 1850. The second potential trigger for phase 3 is based on leaf area index. When the maximum value of leaf area index is reached in phase 2 ([Table 2.36](#)), phase 3 begins. In phase 3, the leaf area index begins to decline in response to a background litterfall rate calculated as the inverse of leaf longevity for the pft as done in the BGC part of the model.

Harvest

Harvest is assumed to occur as soon as the crop reaches maturity. When $GDD_{T_{2m}}$ reaches 100% of GDD_{mat} or the number of days past planting reaches a crop-specific maximum ([Table 2.35](#)), then the crop is harvested. Harvest occurs in one time step using the BGC leaf offset algorithm.

Table 2.35: Crop phenology and morphology parameters for the active crop plant functional types (pfts) in CLM5BGCCROP. Numbers in the first row correspond to the list of pfts in [Table 2.34](#).

	tem-perate corn	spring wheat	temperatuve soybean	cot-ton	rice	sug-ar-cane	tropical corn	tropical soybean
IVT	17, 18	19, 20	23, 24	41, 42	61, 62	67, 68	75, 76	77, 78
$Date_{planting}^{min}$	April 1	April 1	May 1	April 1	Janu- rary 1	Janu- rary 1	March 20	April 15
$Date_{planting}^{max}$	June 15	June 15	June 15	May 31	Febru- ary 28	March 31	April 15	June 31
$T_p(K)$	283.15	280.15	286.15	294.15	294.15	294.15	294.15	294.15
$T_p^{min}(K)$	279.15	272.15	279.15	283.15	283.15	283.15	283.15	283.15
$GDD_{min}(^{\circ}\text{days})$	50	50	50	50	50	50	50	50
base temperature for GDD ($^{\circ}\text{C}$)	8	0	10	10	10	10	10	10
$GDD_{mat}(^{\circ}\text{days})$	950-1850	≤ 1700	≤ 1900	≤ 1700	≤ 2100	950- 1850	950- 1850	≤ 2100
Phase 2 % GDD_{mat}	0.03	0.05	0.03	0.03	0.01	0.03	0.03	0.03
Phase 3 % GDD_{mat}	0.65	0.6	0.5	0.5	0.4	0.65	0.5	0.5
Harvest: days past planting	≤ 165	≤ 150	≤ 150	≤ 160	≤ 150	≤ 300	≤ 160	≤ 150
$z_{top}^{\max}(m)$	2.5	1.2	0.75	1.5	1.8	4	2.5	1
SLA ($\text{m}^2 \text{ leaf g}^{-1} \text{ C}$)	0.05	0.035	0.035	0.035	0.035	0.05	0.05	0.035
χ_L index	-0.5	-0.5	-0.5	-0.5	-0.5	-0.5	-0.5	-0.5
grperc	0.11	0.11	0.11	0.11	0.11	0.11	0.11	0.11
flnr	0.293	0.41	0.41	0.41	0.41	0.293	0.293	0.41
fcur	1	1	1	1	1	1	1	1

Notes: $Date_{planting}^{min}$ and $Date_{planting}^{max}$ are the minimum and maximum planting date in the Northern Hemisphere, the corresponding dates in the Southern Hemisphere apply 6 months later. T_p and T_p^{min} are crop-specific coldest planting temperatures. GDD_{min} is the lowest (for planting) 20-year running mean growing degree-days based on the base temperature threshold in the 7th row, tracked from April to September (NH). GDD_{mat} is a crop's 20-year running mean growing degree-days needed for vegetative and physiological maturity. Harvest occurs at 100% GDD_{mat} or when the days past planting reach the number in the 11th row. Crop growth phases are described in the text. z_{top}^{\max} is the maximum top-of-canopy height of a crop, SLA is specific leaf area. χ_L is the leaf orientation index, equals -1 for vertical, 0 for random, and 1 for horizontal leaf orientation. grperc is the growth respiration factor. flnr is the fraction of leaf N in the Rubisco enzyme. fcur is the fraction of allocation that goes to currently displayed growth.

Allocation

Allocation changes based on the crop phenology phases phenology (section 2.25.2). Simulated C assimilation begins every year upon leaf emergence in phase 2 and ends with harvest at the end of phase 3; therefore, so does the allocation of such C to the crop's leaf, live stem, fine root, and reproductive pools.

Typically, C:N ratios in plant tissue vary throughout the growing season and tend to be lower during early growth stages

and higher in later growth stages. In order to account for this seasonal change, two sets of C:N ratios are established in CLM for the leaf, stem, and fine root of crops: one during the leaf emergence phase (phenology phase 2), and a second during grain fill phase (phenology phase 3). This modified C:N ratio approach accounts for the nitrogen retranslocation that occurs during the grain fill phase (phase 3) of crop growth. Leaf, stem, and root C:N ratios for phase 2 are calculated using the new CLM5 carbon and nitrogen allocation scheme (Chapter 2.19), which provides a target C:N value (Table 2.36) and allows C:N to vary through time. During grain fill (phase 3) of the crop growth cycle, a portion of the nitrogen in the plant tissues is moved to a storage pool to fulfill nitrogen demands of organ (reproductive pool) development, such that the resulting C:N ratio of the plant tissue is reflective of measurements at harvest. All C:N ratios were determined by calibration process, through comparisons of model output versus observations of plant carbon throughout the growing season.

The BGC part of the model keeps track of a term representing excess maintenance respiration, which supplies the carbon required for maintenance respiration during periods of low photosynthesis (Chapter 2.17). Carbon supply for excess maintenance respiration cannot continue to happen after harvest for annual crops, so at harvest the excess respiration pool is turned into a flux that extracts CO₂ directly from the atmosphere. This way any excess maintenance respiration remaining at harvest is eliminated as if such respiration had not taken place.

Leaf emergence

During phase 2, the allocation coefficients (fraction of available C) to each C pool are defined as:

$$\begin{aligned} a_{repr} &= 0 \\ a_{froot} &= a_{froot}^i - (a_{froot}^i - a_{froot}^f) \frac{GDD_{T2m}}{GDD_{mat}} \quad \text{where } \frac{GDD_{T2m}}{GDD_{mat}} \leq 1 \\ a_{leaf} &= (1 - a_{froot}) \cdot \frac{a_{leaf}^i (e^{-b} - e^{-b} \frac{GDD_{T2m}}{h})}{e^{-b} - 1} \quad \text{where } b = 0.1 \\ a_{livestem} &= 1 - a_{repr} - a_{froot} - a_{leaf} \end{aligned} \quad (2.960)$$

where a_{leaf}^i , a_{froot}^i , and a_{froot}^f are initial and final values of these coefficients (Table 2.36), and h is a heat unit threshold defined in section 2.25.2. At a crop-specific maximum leaf area index, L_{max} (Table 2.36), carbon allocation is directed exclusively to the fine roots.

Grain fill

The calculation of a_{froot} remains the same from phase 2 to phase 3. During grain fill (phase 3), other allocation coefficients change to:

$$\begin{aligned} a_{leaf} &= a_{leaf}^{i,3} && \text{when } a_{leaf}^{i,3} \leq a_{leaf}^f \quad \text{else} \\ a_{leaf} &= a_{leaf} \left(1 - \frac{GDD_{T2m} - h}{GDD_{mat} d_L - h}\right)^{d_{alloc}^{leaf}} \geq a_{leaf}^f && \text{where } \frac{GDD_{T2m} - h}{GDD_{mat} d_L - h} \leq 1 \\ a_{livestem} &= a_{livestem}^{i,3} && \text{when } a_{livestem}^{i,3} \leq a_{livestem}^f \quad \text{else} \quad (2.961) \\ a_{livestem} &= a_{livestem} \left(1 - \frac{GDD_{T2m} - h}{GDD_{mat} d_L - h}\right)^{d_{alloc}^{stem}} \geq a_{livestem}^f && \text{where } \frac{GDD_{T2m} - h}{GDD_{mat} d_L - h} \leq 1 \\ a_{repr} &= 1 - a_{froot} - a_{livestem} - a_{leaf} \end{aligned}$$

where $a_{leaf}^{i,3}$ and $a_{livestem}^{i,3}$ (initial values) equal the last a_{leaf} and $a_{livestem}$ calculated in phase 2, d_L , d_{alloc}^{leaf} and d_{alloc}^{stem} are leaf area index and leaf and stem allocation decline factors, and a_{leaf}^f and $a_{livestem}^f$ are final values of these allocation coefficients (Table 2.36).

Nitrogen retranslocation for crops

Nitrogen retranslocation in crops occurs when nitrogen that was used for tissue growth of leaves, stems, and fine roots during the early growth season is remobilized and used for grain development (Pollmer et al. 1979, Crawford et al. 1982, Simpson et al. 1983, Ta and Weiland 1992, Barbottin et al. 2005, Gallais et al. 2006, Gallais et al. 2007). Nitrogen allocation for crops follows that of natural vegetation, is supplied in CLM by the soil mineral nitrogen pool, and depends on C:N ratios for leaves, stems, roots, and organs. Nitrogen demand during organ development is fulfilled through retranslocation from leaves, stems, and roots. Nitrogen retranslocation is initiated at the beginning of the grain fill stage for all crops except soybean, for which retranslocation is after LAI decline. Nitrogen stored in the leaf and stem is moved into a storage retranslocation pool for all crops, and for wheat and rice, nitrogen in roots is also released into the retranslocation storage pool. The quantity of nitrogen mobilized depends on the C:N ratio of the plant tissue, and is calculated as

$$\text{leaf_to_retransn} = N_{leaf} - \frac{C_{leaf}}{CN_{leaf}^f} \quad (2.962)$$

$$\text{stemn_to_retransn} = N_{stem} - \frac{C_{stem}}{CN_{stem}^f} \quad (2.963)$$

$$\text{frootn_to_retransn} = N_{froot} - \frac{C_{froot}}{CN_{froot}^f} \quad (2.964)$$

where C_{leaf} , C_{stem} , and C_{froot} is the carbon in the plant leaf, stem, and fine root, respectively, N_{leaf} , N_{stem} , and N_{froot} is the nitrogen in the plant leaf, stem, and fine root, respectively, and CN_{leaf}^f , CN_{stem}^f , and CN_{froot}^f is the post-grain fill C:N ratio of the leaf, stem, and fine root respectively (Table 2.36). Since C:N measurements are often taken from mature crops, pre-grain development C:N ratios for leaves, stems, and roots in the model are optimized to allow maximum nitrogen accumulation for later use during organ development, and post-grain fill C:N ratios are assigned the same as crop residue. After nitrogen is moved into the retranslocated pool, the nitrogen in this pool is used to meet plant nitrogen demand by assigning the available nitrogen from the retranslocated pool equal to the plant nitrogen demand for each organ ($CN_{[organ]}^f$ in Table 2.36). Once the retranslocation pool is depleted, soil mineral nitrogen pool is used to fulfill plant nitrogen demands.

Harvest

Variables track the flow of grain C and N to food and of all other plant pools, including live stem C and N, to litter. Putting live stem C and N into the litter pool is in contrast to the approach for unmanaged PFTs which puts live stem C and N into dead stem pools first. Leaf and root C and N pools are routed to the litter pools in the same manner as natural vegetation. Whereas food C and N was formerly transferred to the litter pool, CLM5 routes food C and N to a grain product pool where the C and N decay to the atmosphere over one year, similar in structure to the wood product pools. Additionally, CLM5 accounts for the C and N required for crop seeding by removing the seed C and N from the grain product pool during harvest. The crop seed pool is then used to seed crops in the subsequent year. Calculating the crop yields (Equation (2.965)) requires that you sum the GRAINC_TO_FOOD variable for each year, and must account for the proportion of C in the dry crop weight. Here, we assume that grain C is 45% of the total dry weight. Additionally, harvest is not typically 100% efficient, so analysis needs to assume that harvest efficiency is less. We assume a harvest efficiency of 85%.

$$\text{Grain yield(g.m}^{-2}) = \frac{\sum(\text{GRAINC_TO_FOOD}) * 0.85}{0.45} \quad (2.965)$$

Table 2.36: Crop allocation parameters for the active crop plant functional types (pfts) in CLM5BGCCROP. Numbers in the first row correspond to the list of pfts in Table 2.34.

	temperate corn	spring wheat	temperatuue soybean	cotton	rice	sugarcane	tropical corn	tropical soybean
IVT	17, 18	19, 20	23, 24	41, 42	61, 62	67, 68	75, 76	77, 78
a_{leaf}^i	0.8	0.9	0.85	0.85	0.75	0.8	0.8	0.85
L_{max} (m ² m ⁻²)	5	7	6	6	7	5	5	6
a_{froot}^i	0.4	0.1	0.2	0.2	0.1	0.4	0.4	0.2
a_{froot}^f	0.05	0	0.2	0.2	0	0.05	0.05	0.2
a_{leaf}^f	0	0	0	0	0	0	0	0
$a_{livestem}^f$	0	0.05	0.3	0.3	0.05	0	0	0.3
d_L	1.05	1.05	1.05	1.05	1.05	1.05	1.05	1.05
d_{alloc}^{stem}	2	1	5	5	1	2	2	5
d_{alloc}^{leaf}	5	3	2	2	3	5	5	2
CN_{leaf}	25	20	20	20	20	25	25	20
CN_{stem}	50	50	50	50	50	50	50	50
CN_{froot}	42	42	42	42	42	42	42	42
CN_{leaf}^f	65	65	65	65	65	65	65	65
CN_{stem}^f	120	100	130	130	100	120	120	130
CN_{froot}^f	0	40	0	0	40	0	0	0
CN_{grain}	50	50	50	50	50	50	50	50

Notes: Crop growth phases and corresponding variables are described throughout the text. CN_{leaf} , CN_{stem} , and CN_{froot} are the target C:N ratios used during the leaf emergence phase (phase 2).

Other Features

Physical Crop Characteristics

Leaf area index (L) is calculated as a function of specific leaf area (SLA, Table 2.35) and leaf C. Stem area index (S) is equal to $0.1L$ for temperate and tropical corn and sugarcane and $0.2L$ for other crops, as in AgroIBIS. All live C and N pools go to 0 after crop harvest, but the S is kept at 0.25 to simulate a post-harvest “stubble” on the ground.

Crop heights at the top and bottom of the canopy, z_{top} and z_{bot} (m), come from the AgroIBIS formulation:

$$z_{top} = z_{top}^{\max} \left(\frac{L}{L_{\max}-1} \right)^2 \geq 0.05 \text{ where } \frac{L}{L_{\max}-1} \leq 1 \quad (2.966)$$

$$z_{bot} = 0.02 \text{m}$$

where z_{top}^{\max} is the maximum top-of-canopy height of the crop (Table 2.35) and L_{\max} is the maximum leaf area index (Table 2.36).

Interactive Fertilization

CLM simulates fertilization by adding nitrogen directly to the soil mineral nitrogen pool to meet crop nitrogen demands using both industrial fertilizer and manure application. CLM’s separate crop land unit ensures that natural vegetation will not access the fertilizer applied to crops. Fertilizer in CLM5BGCCROP is prescribed by crop functional types and

varies spatially for each year based on the LUMIP land use and land cover change time series (LUH2 for historical and SSPs for future) (Lawrence et al. 2016). One of two fields is used to prescribe industrial fertilizer based on the type of simulation. For non-transient simulations, annual fertilizer application in g N/m²/yr is specified on the land surface data set by the field CONST_FERTNITRO_CFT. In transient simulations, annual fertilizer application is specified on the land use time series file by the field FERTNITRO_CFT, which is also in g N/m²/yr. The values for both of these fields come from the LUMIP time series for each year. In addition to the industrial fertilizer, background manure fertilizer is specified on the parameter file by the field ‘manunitro’. For the current CLM5BGCCROP, manure N is applied at a rate of 0.002 kg N/m²/yr. Because previous versions of CLM (e.g., CLM4) had rapid denitrification rates, fertilizer is applied slowly to minimize N loss (primarily through denitrification) and maximize plant uptake. The current implementation of CLM5 inherits this legacy, although denitrification rates are slower in the current version of the model (Koven et al. 2013). As such, fertilizer application begins during the leaf emergence phase of crop development (phase 2) and continues for 20 days, which helps reduce large losses of nitrogen from leaching and denitrification during the early stage of crop development. The 20-day period is chosen as an optimization to limit fertilizer application to the emergence stage. A fertilizer counter in seconds, f , is set as soon as the leaf emergence phase for crops initiates:

$$f = n \times 86400 \quad (2.967)$$

where n is set to 20 fertilizer application days and 86400 is the number of seconds per day. When the crop enters phase 2 (leaf emergence) of its growth cycle, fertilizer application begins by initializing fertilizer amount to the total fertilizer at each column within the grid cell divided by the initialized f . Fertilizer is applied and f is decremented each time step until a zero balance on the counter is reached.

Biological nitrogen fixation for soybeans

Biological N fixation for soybeans is calculated by the fixation and uptake of nitrogen module (Chapter 2.18) and is the same as N fixation in natural vegetation. Unlike natural vegetation, where a fraction of each pft are N fixers, all soybeans are treated as N fixers.

Latitudinal variation in base growth temperature

For most crops, $GDD_{T_{2m}}$ (growing degree days since planting) is the same in all locations. However, for both rainfed and irrigated spring wheat and sugarcane, the calculation of $GDD_{T_{2m}}$ allows for latitudinal variation:

$$\text{latitudinal variation in base } T = \begin{cases} baset + 12 - 0.4 \times \text{latitude} & 0 \leq \text{latitude} \leq 30 \\ baset + 12 + 0.4 \times \text{latitude} & -30 \leq \text{latitude} \leq 0 \end{cases} \quad (2.968)$$

where $baset$ is the *base temperature for GDD* (7th row) in Table 2.35. Such latitudinal variation in base growth temperature could increase the base temperature, slow down $GDD_{T_{2m}}$ accumulation, and extend the growing season for regions within 30°S to 30°N for spring wheat and sugarcane.

Separate reproductive pool

One notable difference between natural vegetation and crops is the presence of reproductive carbon and nitrogen pools. Accounting for the reproductive pools helps determine whether crops are performing reasonably through yield calculations. The reproductive pool is maintained similarly to the leaf, stem, and fine root pools, but allocation of carbon and nitrogen does not begin until the grain fill stage of crop development. Equation (2.961) describes the carbon and nitrogen allocation coefficients to the reproductive pool. In CLM5BGCCROP, as allocation declines in stem, leaf, and root pools (see section 2.25.2) during the grain fill stage of growth, increasing amounts of carbon and nitrogen are available for grain development.

2.25.3 The irrigation model

The CLM includes the option to irrigate cropland areas that are equipped for irrigation. The application of irrigation responds dynamically to the soil moisture conditions simulated by the CLM. This irrigation algorithm is based loosely on the implementation of [Ozdogan et al. \(2010\)](#).

When irrigation is enabled, the crop areas of each grid cell are divided into irrigated and rainfed fractions according to a dataset of areas equipped for irrigation ([Portmann et al. 2010](#)). Irrigated and rainfed crops are placed on separate soil columns, so that irrigation is only applied to the soil beneath irrigated crops.

In irrigated croplands, a check is made once per day to determine whether irrigation is required on that day. This check is made in the first time step after 6 AM local time. Irrigation is required if crop leaf area > 0 , and the available soil water is below a specified threshold.

The soil moisture deficit D_{irrig} is

$$D_{irrig} = \begin{cases} w_{thresh} - w_{avail} & w_{thresh} > w_{avail} \\ 0 & w_{thresh} \leq w_{avail} \end{cases} \quad (2.969)$$

where w_{thresh} is the irrigation moisture threshold (mm) and w_{avail} is the available moisture (mm). The moisture threshold is

$$w_{thresh} = f_{thresh} (w_{target} - w_{wilt}) + w_{wilt} \quad (2.970)$$

where w_{target} is the irrigation target soil moisture (mm)

$$w_{target} = \sum_{j=1}^{N_{irr}} \theta_{target} \Delta z_j, \quad (2.971)$$

w_{wilt} is the wilting point soil moisture (mm)

$$w_{wilt} = \sum_{j=1}^{N_{irr}} \theta_{wilt} \Delta z_j, \quad (2.972)$$

and f_{thresh} is a tuning parameter. The available moisture in the soil is

$$w_{avail} = \sum_{j=1}^{N_{irr}} \theta_j \Delta z_j, \quad (2.973)$$

N_{irr} is the index of the soil layer corresponding to a specified depth z_{irrig} ([Table 2.37](#)) and Δz_j is the thickness of the soil layer in layer j (section 2.2.2). θ_j is the volumetric soil moisture in layer j (section 2.7.3). θ_{target} and θ_{wilt} are the target and wilting point volumetric soil moisture values, respectively, and are determined by inverting (2.410) using soil matric potential parameters Ψ_{target} and Ψ_{wilt} ([Table 2.37](#)). After the soil moisture deficit D_{irrig} is calculated, irrigation in an amount equal to $\frac{D_{irrig}}{T_{irrig}}$ (mm/s) is applied uniformly over the irrigation period T_{irrig} (s). Irrigation water is applied directly to the ground surface, bypassing canopy interception (i.e., added to $q_{grnd,liq}$: section 2.7.1).

To conserve mass, irrigation is removed from river water storage ([Chapter 2.14](#)). When river water storage is inadequate to meet irrigation demand, there are two options: 1) the additional water can be removed from the ocean model, or 2) the irrigation demand can be reduced such that river water storage is maintained above a specified threshold.

Table 2.37: Irrigation parameters

Parameter	
f_{thresh}	1.0
z_{irrig} (m)	0.6
Ψ_{target} (mm)	-3400
Ψ_{wilt} (mm)	-150000

2.26 Transient Land Use and Land Cover Change

CLM includes a treatment of mass and energy fluxes associated with prescribed temporal land use and land cover change (LULCC). The model uses an annual time series of the spatial distribution of the natural and crop land units of each grid cell, in combination with the distribution of PFTs and CFTs that exist in those land units. Additional land use is prescribed through annual crop specific management of nitrogen fertilizer and irrigation described further in Chapter 25, and through wood harvest on tree PFTs. For changes in the distributions of natural and crop vegetation CLM diagnoses the change in area of the PFTs and CFTs on January 1st of each model year and then performs mass and energy balance accounting necessary to represent the expansion and contraction of the PFT and CFT areas. The biogeophysical impacts of LULCC are simulated through changes in surface properties which in turn impact the surface albedo, hydrology, and roughness which then impact fluxes of energy, moisture and momentum to the atmosphere under the altered properties. Additionally changes in energy and moisture associated with changes in the natural and crop vegetation distribution are accounted for through small fluxes to the atmosphere. The biogeochemical impacts of LULCC are simulated through changes in CLM carbon pools and fluxes as shown in Figure xx.x and described further in Chapter 16.

2.26.1 Annual Transient Land Use and Land Cover Data and Time Interpolation

The changes in area over time associated with changes in natural and crop vegetation and the land use on that vegetation are prescribed through a forcing dataset, referred to here as the *landuse.timeseries* dataset. The *landuse.timeseries* dataset consists of an annual time series of global grids, where each annual time slice describes the fractional area occupied by all PFTs and CFTs along with the nitrogen fertilizer and irrigation fraction of each crop CFT, and the annual wood harvest applied to tree PFTs. Changes in area of PFTs and CFTs are performed annually on the first time step of January 1st of the year. Fertilizer application, irrigation and wood harvest for each PFT and CFT are performed at each model time step depending on rules from the crop and natural vegetation phenology models. The irrigation fraction is set annually however fertilizer application and wood harvest are set from a time-interpolation of the application rates from the two bracketing annual time slices in the *landuse.timeseries* dataset.

As a special case, when the time dimension of the *landuse.timeseries* dataset starts at a later year than the current model time step, the first time slice from the *landuse.timeseries* dataset is used to represent the current time step PFT and CFT fractional area distributions. Similarly, when the time dimension of the *landuse.timeseries* dataset stops at an earlier year than the current model time step, the last time slice of the *landuse.timeseries* dataset is used. Thus, the simulation will have invariant representations of PFT and CFT distributions through time for the periods prior to and following the time duration of the *landuse.timeseries* dataset, with transient PFT and CFT distributions during the period covered by the *landuse.timeseries* dataset.

The following equations capture this logic, where $year_{cur}$ is the calendar year for the current timestep, $landuse.timeseries_year(1)$ and $landuse.timeseries_year(nyears)$ are the first and last calendar years in the *landuse.timeseries* dataset, respectively, $nyears$ is the number of years in the *landuse.timeseries* dataset, nt_1 and nt_2 are the two bracketing years used in the interpolation algorithm, and n is the index value for the $landuse.timeseries_year$ array corresponding to $landuse.timeseries_year(n) = year_{cur}$:

$$nt_1 = \begin{cases} 1 & \text{for } year_{cur} < landuse.timeseries_year(1) \\ n & \text{for } landuse.timeseries_year(1) \leq year_{cur} < landuse.timeseries_year(nyears) \\ nyears & \text{for } year_{cur} \geq landuse.timeseries_year(nyears) \end{cases} \quad (2.974)$$

$$nt_2 = \begin{cases} 1 & \text{for } year_{cur} < landuse.timeseries_year(1) \\ n + 1 & \text{for } landuse.timeseries_year(1) \leq year_{cur} < landuse.timeseries_year(nyears) \\ nyears & \text{for } year_{cur} \geq landuse.timeseries_year(nyears) \end{cases} \quad (2.975)$$

Interpolation of fertilizer and wood harvest rates between annual time slices in the *landuse.timeseries* dataset uses a simple linear algorithm, based on the conversion of the current time step information into a floating-point value for the

number of calendar days since January 1 of the current model year ($cday$). The interpolation weight for the current time step tw_{cday} is

$$tw_{cday} = \frac{366 - cday}{365} \quad (2.976)$$

where the numerator is 366 instead of 365 because the time manager function for CLM returns a value of $cday = 1.0$ for midnight Greenwich mean time on January 1. With weights $w_p(nt_1)$ and $w_p(nt_2)$ obtained from the *landuse.timeseries* dataset for fertilizer and wood harvest p at the bracketing annual time slices nt_1 and nt_2 , the interpolated application rate for the current time step ($w_{p,t}$) is

$$w_{p,t} = tw_{cday} [w_p(nt_1) - w_p(nt_2)] + w_p(nt_2) \quad (2.977)$$

The form of this equation is designed to improve roundoff accuracy performance, and guarantees $w_{p,t}$ stays in the range $[0,1]$. Note that values for $w_p(nt_1)$, $w_p(nt_2)$, and $w_{p,t}$ are fractional weights at the column level of the subgrid hierarchy.

The change in weight for a fertilizer or wood harvest rate between the current and previous time steps (Δw_p) is

$$\Delta w_p = w_p^n - w_p^{n-1} \quad (2.978)$$

where n denotes the current time step. The rate of application increases for $\Delta w_p > 0$ and decreases for $\Delta w_p < 0$.

2.26.2 Mass and Energy Conservation

Mass conservation is maintained across PFT and CFT weight transitions by summing up all the carbon, nitrogen, water and energy state variables to get the total vegetated land units value before ($W_{tot,1}$) and after ($W_{tot,2}$) the new PFT and CFT weights are calculated. Transitions are performed on above ground variables first and then at the land unit level for below ground variables second. For example the hydrological balance is calculated, $W_{tot,1}$ is

$$W_{tot,1} = W_a + W_{sno} + \sum_{i=1}^{N_{levgrnd}} (w_{liq,i} + w_{ice,i}) + \sum_{j=1}^{npft} (W_{can,j} wt_{j,1}) \quad (2.979)$$

where W_a is the aquifer water, W_{sno} is the snow water, $w_{liq,i}$ and $w_{ice,i}$ are the liquid and ice soil water contents, $W_{can,j}$ is the canopy water content for PFT and CFT j , and $wt_{j,1}$ is the PFT or CFT weight for j . For the situation where PFT and CFT weights are changing, any difference between $W_{tot,1}$ and $W_{tot,2}$ are due to differences in the total canopy water before and after the PFT and CFT weight change. To ensure conservation, the typically very small difference between $W_{tot,2}$ and $W_{tot,1}$ is subtracted from the grid cell runoff

$$R_{liq} = R_{liq} + W_{tot,2} - W_{tot,1}. \quad (2.980)$$

Total energy is unperturbed in this case and therefore an energy conservation treatment is not required. Changing the area of natural and crop land units in association with the change in PFTs and CFTs results in changes in the soil/snow columns and land unit area. To address these additional changes, conservation of mass and energy among the soil/snow columns and land units is performed as a secondary calculation once all above ground PFT and CFT changes have been done.

2.26.3 Annual Transient Land Cover Dataset Development

This section describes the development of the *landuse.timeseries* dataset. Development of this dataset involves the translation of harmonized datasets of LULCC for the historical period and for the different Shared Socioeconomic Pathway (SSP) - Representative Concentration Pathway (RCP) scenarios. Additionally, LULCC time series are to be generated for the Last Millennium and the extension beyond 2100 experiments of CMIP6.

LUH2 Transient Land Use and Land Cover Change Dataset

To coordinate the processing and consistency of LULCC data between the historical period (1850-2015) and the six SSP-RCP (2016-2100) scenarios derived from Integrated Assessment Models (IAM), the University of Maryland and the University of New Hampshire research groups (Louise Chini, George Hurtt, Steve Frolking and Ritvik Sahajpal; luh.umd.edu) produced a new version of the Land Use Harmonized version 2 (LUH2) transient datasets for use with Earth System Model simulations. The new data sets are the product of the Land Use Model Intercomparison Project (LUMIP; <https://cmip.ucar.edu/lumip>) as part of the Coupled Model Intercomparison Project 6 (CMIP6). The historical component of the transient LULCC dataset has agriculture and urban land use based on HYDE 3.2 with wood harvest based on FAO, Landsat and other sources, for the period 850-2015. The SSP-RCP transient LULCC components (2015-2100) are referred to as the LUH2 Future Scenario datasets. The LULCC information is provided at 0.25 degree grid resolution and includes fractional grid cell coverage by the 12 land units of:

Primary Forest, Secondary Forest, Primary Non-Forest, Secondary Non-Forest,

Pasture, Rangeland, Urban,

C3 Annual Crop, C4 Annual Crop, C3 Perennial Crop, C4 Perennial Crop, and C3 Nitrogen Fixing Crop.

The new land unit format is an improvement on the CMIP5 LULCC datasets as they: provide Forest and Non Forest information in combination with Primary and Secondary land; differentiate between Pasture and Rangelands for grazing livestock; and specify annual details on the types of Crops grown and management practices applied in each grid cell. Like the CMIP5 LULCC datasets Primary vegetation represents the fractional area of a grid cell with vegetation undisturbed by human activities. Secondary vegetation represents vegetated areas that have recovered from some human disturbance; this could include re-vegetation of pasture and crop areas as well as primary vegetation areas that have been logged. In this manner the land units can change through deforestation from Forested to Non Forested land and in the opposite direction from Non Forested to Forested land through reforestation or afforestation without going through the Crop, Pasture or Rangeland states.

The LUH2 dataset provides a time series of land cover states as well as a transition matrices that describes the annual fraction of land that is transformed from one land unit category to another (e.g. Primary Forest to C3 Annual Crop, Pasture to C3 Perennial Crop, etc.; Lawrence et al. 2016). Included in these transition matrices is the total conversion of one land cover type to another referred to as Gross LULCC. This value can be larger than the sum of the changes in the state of a land unit from one time period to the next known as the Net LULCC. This difference is possible as land unit changes can occur both from the land unit and to the land unit at the same time. An example of this difference occurs with shifting cultivation where Secondary Forest can be converted to C3 Annual Crop at the same time as C3 Annual Crop is abandoned to Secondary Forest.

The transition matrices also provide harmonized prescriptions of wood harvest both in area of the grid cell harvested and in the amount of biomass carbon harvested. The wood harvest biomass amount includes a 30% slash component inline with the CMIP5 LULCC data described in (Hurtt et al. 2011). The harvest area and carbon amounts are prescribed for the five classes of: Primary Forest, Primary Non-Forest, Secondary Mature Forest, Secondary Young Forest, and Secondary Non-Forest.

Additional land use management is prescribed on the Crop land units for nitrogen fertilization and irrigation equipped land. The fertilizer application and the irrigation fraction is prescribed for each Crop land unit in a grid cell individually for each year of the time series. The wood harvest and crop management are both prescribed spatially on the same 0.25 degree grid as the land use class transitions.

Representing LUH2 Land Use and Land Cover Change in CLM5

To represent the LUH2 transient LULCC dataset in CLM5, the annual fractional composition of the twelve land units specified in the dataset needs to be faithfully represented with a corresponding PFT and CFT mosaics of CLM. CLM5 represents the land surface as a hierarchy of sub-grid types: glacier; lake; wetland; urban; vegetated land; and crop land. The vegetated land is further divided into a mosaic of Plant Functional Types (PFTs), while the crop land is divided into a mosaic of Crop Functional Types (CFTs).

To support this translation task the CLM5 Land Use Data tool has been built that extends the methods described in Lawrence et al (2012) to include all the new functionality of CMIP6 and CLM5 LULCC. The tool translates each of the LUH2 land units for a given year into fractional PFT and CFT values based on the current day CLM5 data for the land unit in that grid cell. The current day land unit descriptions are generated from 1km resolution MODIS, MIRCA2000, ICESAT, AVHRR, SRTM, and CRU climate data products combined with reference year LUH2 land unit data, usually set to 2005. Where the land unit does not exist in a grid cell for the current day, the land unit description is generated from nearest neighbors with an inverse distance weighted search algorithm.

The Land Use Data tool produces raw vegetation, crop, and management data files which are combined with other raw land surface data to produce the CLM5 initial surface dataset and the dynamic *landuse.timeseries* dataset with the CLM5 mksurfdata_map tool. The schematic of this entire process from LUH2 time series and high resolution current day data to the output of CLM5 surface datasets from the mksurfdata_map tool is shown in Figure 21.2.

The methodology for creating the CLM5 transient PFT and CFT dataset is based on four steps which are applied across all of the historical and future time series. The first step involves generating the current day descriptions of natural and managed vegetation PFTs at 1km resolution from the global source datasets, and the current day description of crop CFTs at the 10km resolution from the MIRCA 2000 datasets. The second step combines the current day (2005) LUH2 land units with the current day CLM5 PFT and CFT distributions to get CLM5 land unit descriptions in either PFTs or CFTs at the LUH2 resolution of 0.25 degrees. The third step involves combining the LUH2 land unit time series with the CLM5 PFT and CFT descriptions for that land unit to generate the CLM5 raw PFT and CFT time series in the *landuse.timeseries* file. At this point in the process management information in terms of fertilizer, irrigation and wood harvest are added to the CLM5 PFT and CFT data to complete the CLM5 raw PFT and CFT files. The final step is to combine these files with the other raw CLM5 surface data files in the mksurfdata_map tool.

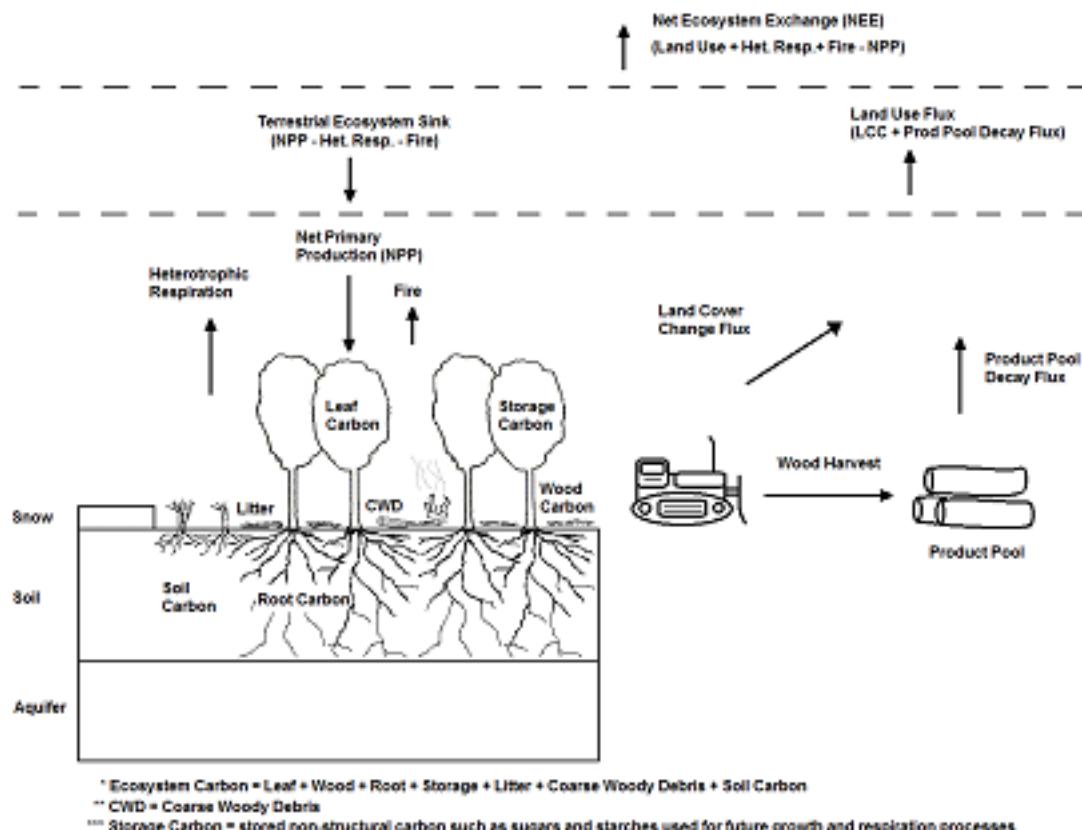


Figure 2.20: Schematic of land cover change impacts on CLM carbon pools and fluxes.

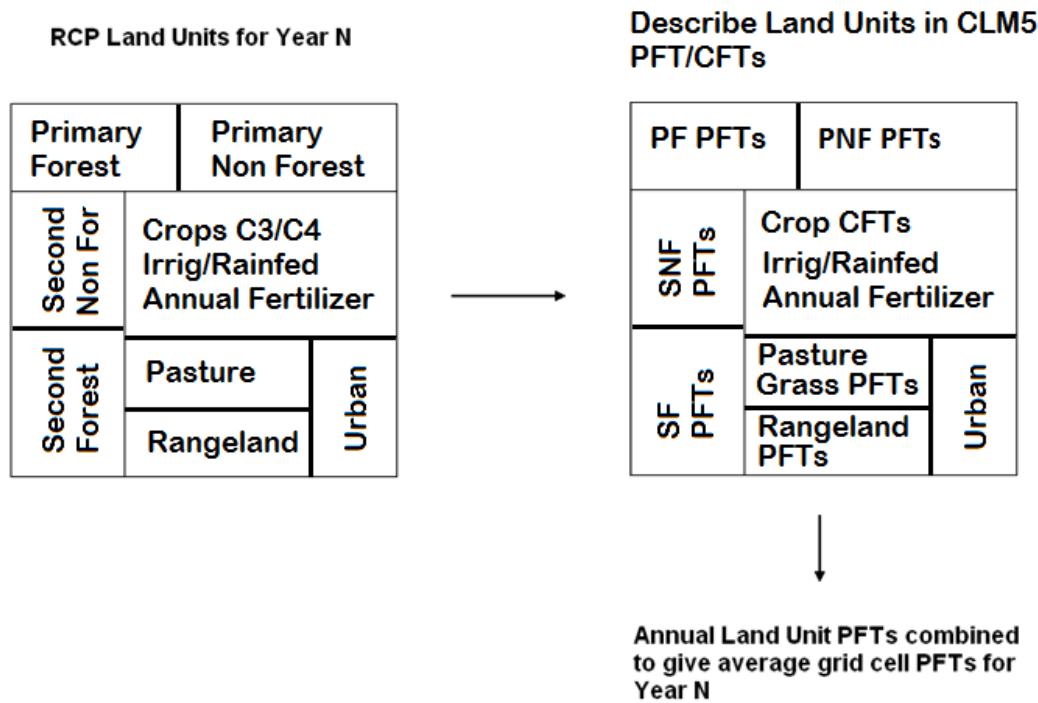


Figure 2.21: Schematic of translation of annual LUH2 land units to CLM5 plant and crop functional types.

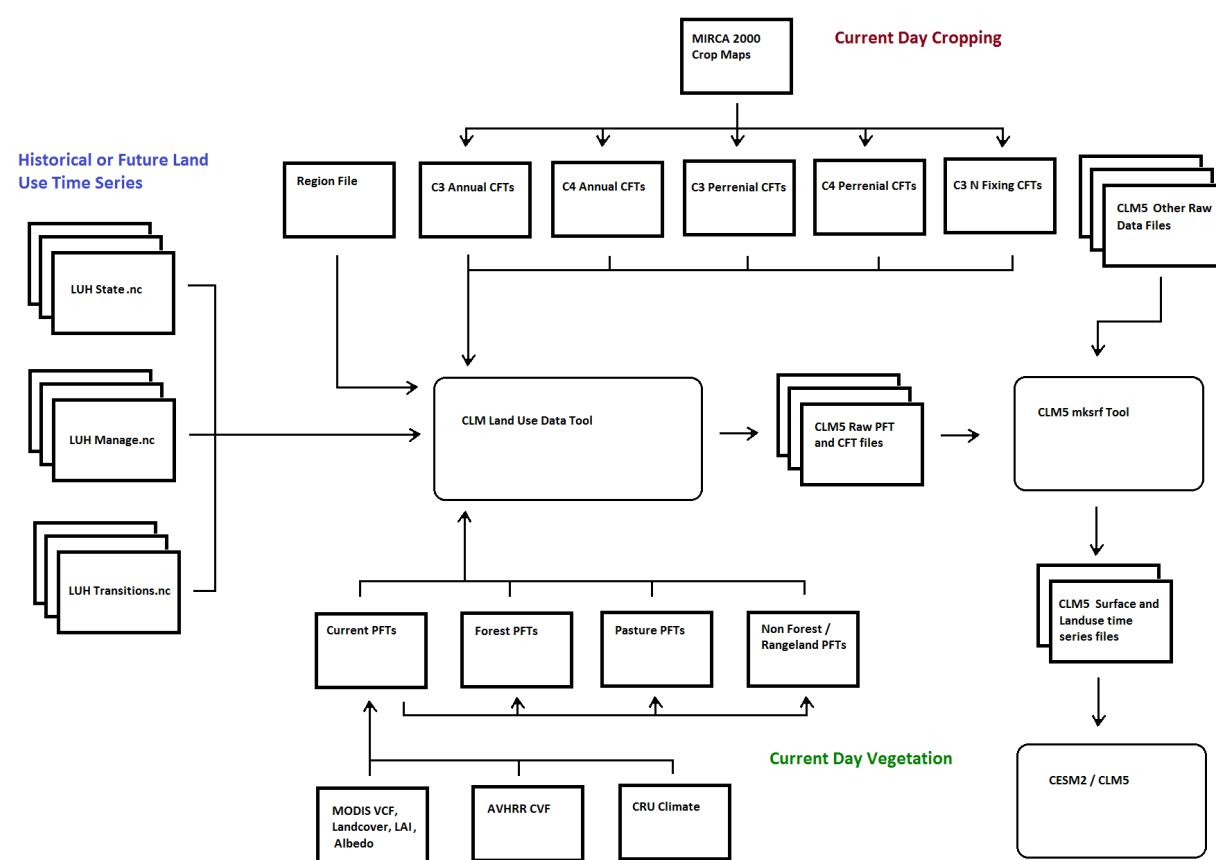


Figure 2.22: Workflow of CLM5 Land Use Data Tool and Mksurfdata_map Tool

2.27 Dynamic Global Vegetation

2.27.1 What has changed

- Deprecation of the dynamic global vegetation model (DGVM): The CLM5.0 model contains the legacy ‘CNDV’ code, which runs the CLM4(CN) model in combination with the LPJ-derived dynamics vegetation model introduced in CLM3. While this capacity has not technically been removed from the model, the DGVM has not been tested in the development of CLM5 and is no longer scientifically supported.
- Introduction of FATES: The Functionally Assembled Terrestrial Ecosystem Simulator (FATES) is the actively developed DGVM for the CLM5. See

2.28 Technical Documentation for FATES

FATES is the “Functionally Assembled Terrestrial Ecosystem Simulator”. It is an external module which can run within a given “Host Land Model” (HLM). Currently (November 2017) implementations are supported in both the Community Land Model(CLM) and in the land model of the E3SM Dept. of Energy Earth System Model.

FATES was derived from the CLM Ecosystem Demography model (CLM(ED)), which was documented in:

Fisher RA, Muszala S, Verteinstein M, Lawrence P, Xu C, McDowell NG, Knox RG, Koven C, Holm J, Rogers BM, Lawrence D. Taking off the training wheels: the properties of a dynamic vegetation model without climate envelopes. Geoscientific Model Development Discussions. 2015 Apr 1;8(4).

and this technical note was first published as an appendix to that paper.

<https://pdfs.semanticscholar.org/396c/b9f172cb681421ed78325a2237bfb428eece.pdf>

2.28.1 Introduction

The Ecosystem Demography (‘ED’), concept within FATES is derived from the work of *Moorcroft et al. (2001)* and is a cohort model of vegetation competition and co-existence, allowing a representation of the biosphere which accounts for the division of the land surface into successional stages, and for competition for light between height structured cohorts of representative trees of various plant functional types.

The implementation of the Ecosystem Demography concept within FATES links the surface flux and canopy physiology concepts in the CLM/E3SM with numerous additional developments necessary to accommodate the new model also documented here. These include a version of the SPITFIRE (Spread and InTensity of Fire) model of *Thonicke et al. (2010)*, and an adoption of the concept of *Perfect Plasticity Approximation* approach of *Purves et al. 2008, Lichstein et al. 2011* and *Weng et al. 2014*, in accounting for the spatial arrangement of crowns. Novel algorithms accounting for the fragmentation of coarse woody debris into chemical litter streams, for the physiological optimisation of canopy thickness, for the accumulation of seeds in the seed bank, for multi-layer multi-PFT radiation transfer, for drought deciduous and cold deciduous phenology, for carbon storage allocation, and for tree mortality under carbon stress, are also included and presented here.

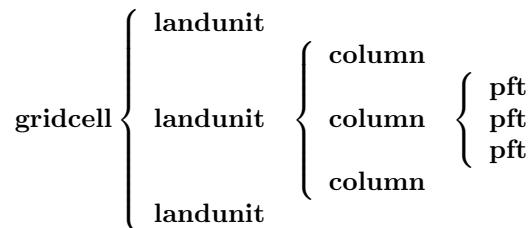
Numerous other implementations of the Ecosystem Demography concept exist (See Fisher et al. 2017 for a review of these) Therefore, to avoid confusion between the concept of ‘Ecosystem Demography’ and the implementation of this concept in different models, the CLM(ED) implementation described by Fisher et al. (2015) will hereafter be called ‘FATES’ (the Functionally Assembled Terrestrial Ecosystem Simulator).

2.28.2 The representation of ecosystem heterogeneity in FATES

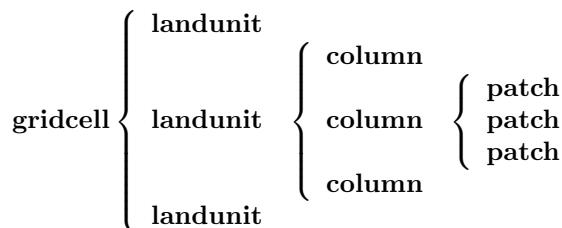
The terrestrial surface of the Earth is heterogeneous for many reasons, driven by variations in climate, edaphic history, ecological variability, geological forcing and human interventions. Land surface models represent this variability first by introducing a grid structure to the land surface, allowing different atmospheric forcings to operate in each grid cell, and subsequently by representing ‘sub-grid’ variability in the surface properties. In the CLM, the land surface is divided into numerous ‘landunits’ corresponding to the underlying condition of the surface (e.g. soils, ice, lakes, bare ground) and then ‘columns’ referring to elements of the surface that share below ground resources (water & nutrients). Within the soil landunit, for example, there are separate columns for crops, and for natural vegetation, as these are assumed to use separate resource pools. The FATES model at present only operates on the naturally vegetated column. The soil column is sub-divided into numerous tiles, that correspond to statistical fractions of the potentially vegetated land area. In the CLM 4.5 (and all previous versions of the model), sub-grid tiling operates on the basis of plant functional types (PFTs). That is, each piece of land is assumed to be occupied by only one plant functional type, with multiple PFT-specific tiles sharing a common soil water and nutrient pool. This PFT-based tiling structure is the standard method used by most land surface models deployed in climate prediction.

The introduction of the Ecosystem Demography concept introduces significant alterations to the representation of the land surface in the CLM. In FATES, the tiling structure represents the disturbance history of the ecosystem. Thus, some fraction of the land surface is characterized as ‘recently disturbed’, some fraction has escaped disturbance for a long time, and other areas will have intermediate disturbances. Thus the ED concept essentially discretizes the trajectory of succession from disturbed ground to ‘mature’ ecosystems. Within FATES, each “disturbance history class” is referred to as a ‘patch’. The word “patch” has many possible interpretations, so it is important to note that: **there is no spatial location associated with the concept of a ‘patch’ . It refers to a fraction of the potential vegetated area consisting of all parts of the ecosystem with similar disturbance history.**

The ‘patch’ organizational structure in CLM thus replaces the previous ‘PFT’ structure in the organization heirarchy. The original hierarchical land surface organizational structure of CLM as described in *Oleson et al. 2013* may be depicted as:



and the new structure is altered to the following:



Thus, each gridcell becomes a matrix of ‘patches’ that are conceptualized by their ‘age since disturbance’ in years. This is the equivalent of grouping together all those areas of a gridcell that are ‘canopy gaps’, into a single entity, and all those areas that are ‘mature forest’ into a single entity.

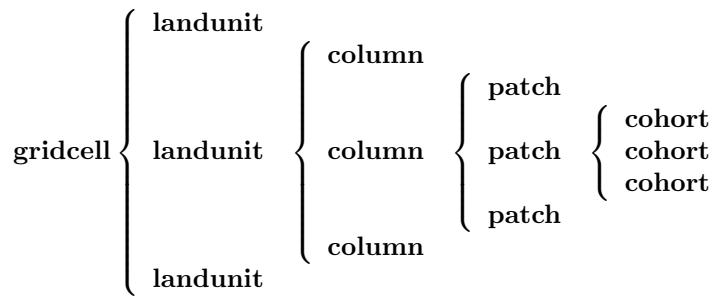
Cohortized representation of tree populations

Each common-disturbance-history patch is a notional ecosystem that might in reality contain numerous individual plants which vary in their physiological attributes, in height and in spatial position. One way of addressing this

heterogeneity is to simulate a forest of specific individuals, and to monitor their behavior through time. This is the approach taken by “gap” and individual-based models (*Smith et al. 2001, Sato et al. 2007, Uriarte et al. 2009, Fyllas et al. 2014*). The depiction of individuals typically implies that the outcome of the model is stochastic. This is because we lack the necessary detailed knowledge to simulate the individual plant’s fates. Thus gap models imply both stochastic locations and mortality of plants. Thus, (with a genuinely random seed) each model outcome is different, and an ensemble of model runs is required to generate an average representative solution. Because the random death of large individual trees can cause significant deviations from the mean trajectory for a small plot (a typical simulated plot size is 30m x 30 m) the number of runs required to minimize these deviations is large and computationally expensive. For this reason, models that resolve individual trees typically use a physiological timestep of one day or longer (e.g. *Smith et al. 2001, Xiaodong et al. 2005, Sato et al. 2007*)

The approach introduced by the Ecosystem Demography model *Moorcroft et al. 2001* is to group the hypothetical population of plants into “cohorts”. In the notional ecosystem, after the land-surface is divided into common-disturbance-history patches, the population in each patch is divided first into plant functional types (the standard approach to representing plant diversity in large scale vegetation models), and then each plant type is represented as numerous height classes. Importantly, **for each PFT/height class bin, we model *one* representative individual plant, which tracks the average properties of this ‘cohort’ of individual plants.** Thus, each common-disturbance-history patch is typically occupied by a set of cohorts of different plant functional types, and different height classes within those plant functional types. Each cohort is associated with a number of identical trees, n_{coh} (where coh denotes the identification or index number for a given cohort)..

The complete hierarchy of elements in FATES is therefore now described as follows:



Discretization of cohorts and patches

Newly disturbed land and newly recruited seedlings can in theory be generated at each new model timestep as the result of germination and disturbance processes. If the new patches and cohorts established at *every* timestep were tracked by the model structure, the computational load would of course be extremely high (and thus equivalent to an individual-based approach). A signature feature of the ED model is the system by which *functionally equivalent* patches and cohorts are fused into single model entities to save memory and computational time.

¹ This functionality requires that criteria are established for the meaning of *functional equivalence*, which are by necessity slightly subjective, as they represent ways of abstracting reality into a more tractable mathematical representation. As an example of this, for height-structured cohorts, we calculate the relativized differences in height (h_{coh} , m) between two cohorts of the same pft, p and q as

$$d_{hite,p,q} = \frac{\text{abs.}(h_p - h_q)}{\frac{1}{2}(h_p + h_q)}$$

If $d_{hite,p,q}$ is smaller than some threshold t_{ch} , and they are of the same plant functional type, the two cohorts are considered equivalent and merged to form a third cohort r , with the properties of cohort p and q averaged such that they conserve mass. The model parameter t_{ch} can be adjusted to adjust the trade-off between simulation accuracy and computational load. There is no theoretical optimal value for this threshold but it may be altered to have finer or coarser model resolutions as needed.

¹ This description covers algorithms in the ‘fuse_cohorts’ subroutine.

² Similarly, for common-disturbance-history patches, we again assign a threshold criteria, which is then compared to the difference between patches m and n , and if the difference is less than some threshold value (t_p) then patches are merged together, otherwise they are kept separate. However, in contrast with height-structured cohorts, where the meaning of the difference criteria is relatively clear, how the landscape should be divided into common-disturbance-history units is less clear. Several alternative criteria are possible, including Leaf Area Index, total biomass and total stem basal area.

In this implementation of FATES we assess the amount of above-ground biomass in each PFT/plant diameter bin. Biomass is first grouped into fixed diameter bins for each PFT (ft) and a significant difference in any bin will cause patches to remain separated. This means that if two patches have similar total biomass, but differ in the distribution of that biomass between diameter classes or plant types, they remain as separate entities. Thus

$$B_{profile,m,dc,ft} = \sum_{d_{c,min}}^{d_{c,max}} (B_{ag,coh} n_{coh})$$

$B_{profile,m,dc,ft}$ is the binned above-ground biomass profile for patch m . d_c is the diameter class. $d_{c,min}$ and $d_{c,max}$ are the lower and upper boundaries for the d_c diameter class. $B_{ag,coh}$ and n_{coh} depict the biomass (KgC m^{-2}) and the number of individuals of each cohort respectively. A difference matrix between patches m and n is thus calculated as

$$d_{biomass,mn,dc,ft} = \frac{\text{abs}(B_{profile,m,hc,ft} - B_{profile,n,hc,ft})}{\frac{1}{2}(B_{profile,m,hc,ft} + B_{profile,n,hc,ft})}$$

If all the values of $d_{biomass,mn,dc,ft}$ are smaller than the threshold, t_p , then the patches m and n are fused together to form a new patch o .

To increase computational efficiency and to simplify the coding structure of the model, the maximum number of patches is capped at $P_{no,max}$. To force the fusion of patches down to this number, the simulation begins with a relatively sensitive discretization of patches ($t_p = 0.2$) but if the patch number exceeds the maximum, the fusion routine is repeated iteratively until the two most similar patches reach their fusion threshold. This approach maintains an even discretization along the biomass gradient, in contrast to, for example, simply fusing the oldest or youngest patches together.

³ The area of the new patch ($A_{patch,o}$, m^2) is the sum of the area of the two existing patches,

$$A_{patch,o} = A_{patch,n} + A_{patch,m}$$

and the cohorts ‘belonging’ to patches m and n now co-occupy patch o . The state properties of m and n (litter, seed pools, etc.) are also averaged in accordance with mass conservation .

Linked Lists: the general code structure of FATES

⁴ The number of patches in each natural vegetation column and the number of cohorts in any given patch are variable through time because they are re-calculated for each daily timestep of the model. The more complex an ecosystem, the larger the number of patches and cohorts. For a slowly growing ecosystem, where maximum cohort size achieved between disturbance intervals is low, the number of cohorts is also low. For fast-growing ecosystems where many plant types are viable and maximum heights are large, more cohorts are required to represent the ecosystem with adequate complexity.

In terms of variable structure, the creation of an array whose size could accommodate every possible cohort would mean defining the maximum potential number of cohorts for every potential patch, which would result in very large amounts of wasted allocated memory, on account of the heterogeneity in the number of cohorts between complex and simple ecosystems (n.b. this does still happen for some variables at restart timesteps). To resolve this, the cohort

² This description covers algorithms in the ‘fuse_patches’ subroutine.

³ This description covers algorithms in the ‘fuse_2_patches’ subroutine.

⁴ This description covers the structure of code in all modules in clm4_5 that are located in ‘ED’ subdirectories

structure in FATES model does not use an array system for internal calculations. Instead it uses a system of *linked lists* where each cohort structure is linked to the cohorts larger than and smaller than itself using a system of pointers. The shortest cohort in each patch has a ‘shorter’ pointer that points to the *null* value, and the tallest cohort has a ‘taller’ pointer that points to the null value.

Instead of iterating along a vector indexed by *coh*, the code structures typically begin at the tallest cohort in a given patch, and iterate until a null pointer is encountered.

Using this structure, it is therefore possible to have an unbounded upper limit on cohort number, and also to easily alter the ordering of cohorts if, for example, a cohort of one functional type begins to grow faster than a competitor of another functional type, and the cohort list can easily be re-ordered by altering the pointer structure. Each cohort has *pointers* indicating to which patch and gridcell it belongs. The patch system is analogous to the cohort system, except that patches are ordered in terms of their relative age, with pointers to older and younger patches where *cp₁* is the oldest:

Indices used in FATES

Some of the indices used in FATES are similar to those used in the standard CLM4.5 model; column (*c*), land unit(*l*), grid cell(*g*) and soil layer (*j*). On account of the additional complexity of the new representation of plant function, several additional indices are introduced that describe the discritization of plant type, fuel type, litter type, plant height, canopy identity, leaf vertical structure and fuel moisture characteristics. To provide a reference with which to interpret the equations that follow, they are listed here.

Table 2.38: Table of subscripts used in this document

Parameter Symbol	Parameter Name
<i>ft</i>	Plant Functional Type
<i>fc</i>	Fuel Class
<i>lsc</i>	Litter Size Class
<i>coh</i>	Cohort Index
<i>patch</i>	Patch Index
<i>Cl</i>	Canopy Layer
<i>z</i>	Leaf Layer
<i>mc</i>	Moisture Class

Cohort State Variables

The unit of allometry in the ED model is the cohort. Each cohort represents a group of plants with similar functional types and heights that occupy portions of column with similar disturbance histories. The state variables of each cohort therefore consist of several pieces of information that fully describe the growth status of the plant and its position in the ecosystem structure, and from which the model can be restarted. The state variables of a cohort are as follows:

Table 2.39: State Variables of ‘cohort’ structure

Quantity	Variable name	Units	Notes
Plant Functional Type	ft_{coh}	integer	
Number of Individuals	n_{coh}	n per 10000m ⁻²	
Height	h_{coh}	m	
Diameter	dbh_{coh}	cm	
Structural Biomass	$b_{struc,coh}$	KgC plant ⁻¹	Stem wood (above and below ground)
Alive Biomass	$b_{alive,coh}$	KgC plant ⁻¹	Leaf, fine root and sapwood
Stored Biomass	$b_{store,coh}$	KgC plant ⁻¹	Labile carbon reserve
Leaf memory	$l_{memory,coh}$	KgC plant ⁻¹	Leaf mass when leaves are dropped
Canopy Layer	$C_{l,coh}$	integer	1 = top layer
Phenological Status	$S_{phen,coh}$	integer	1=leaves off. 2=leaves on
Canopy trimming	$C_{trim,coh}$	fraction	1.0=max leaf area
Patch Index	p_{coh}	integer	To which patch does this cohort belong?

Patch State Variables

A patch, as discuss earlier, is a fraction of the landscape which contains ecosystems with similar structure and disturbance history. A patch has no spatial location. The state variables, which are ‘ecosystem’ rather than ‘tree’ scale properties, from which the model can be restarted, are as follows

Table 2.40: State variables of ‘patch’ structure

Quantity	Variable name	Units	Indexed By
Area	A_{patch}	m ²	•
Age	age_{patch}	years	•
Seed	$seed_{patch}$	KgC m ⁻²	ft
Leaf Litter	$l_{litter,patch}$	KgC m ⁻²	ft
Root Litter	$r_{litter,patch}$	KgC m ⁻²	ft
AG Coarse Woody Debris	:math: '{CWD}_{A,patch}'	KgC m ⁻²	Size Class (lsc)
BG Coarse Woody Debris	:math: '{CWD}_{B,patch}'	KgC m ⁻²	Size Class (lsc)
Canopy Spread	$S_{c,patch}$	•	Canopy Layer
Column Index	l_{patch}	integer	•

Model Structure

Code concerned with the Ecosystem Demography model interfaces with the CLM model in four ways: i) During initialization, ii) During the calculation of surface processes (albedo, radiation absorption, canopy fluxes) each model time step (typically half-hourly), iii) During the main invocation of the ED model code at the end of each day. Daily cohort-level NPP is used to grow plants and alter the cohort structures, disturbance processes (fire and mortality) operate to alter the patch structures, and all fragmenting carbon pool dynamics are calculated. iv) during restart

reading and writing. The net assimilation (NPP) fluxes attributed to each cohort are accumulated throughout each daily cycle and passed into the ED code as the major driver of vegetation dynamics.

2.28.3 Initialization of vegetation from bare ground

[5]_If the model is restarted from a bare ground state (as opposed to a pre-existing vegetation state), the state variables above are initialized as follows. First, the number of plants per PFT is allocated according to the initial seeding density (S_{init} , individuals per m²) and the area of the patch A_{patch} , which in the first timestep is the same as the area of the notional ecosystem A_{tot} . The model has no meaningful spatial dimension, but we assign a notional area such that the values of ' n_{coh} ' can be attributed. The default value of A_{tot} is one hectare (10,000 m²), but the model will behave identically irrepective of the value of this parameter.

$$n_{coh,0} = S_{init} A_{patch}$$

Each cohort is initialized at the minimum canopy height $h_{min,ft}$, which is specified as a parameter for each plant functional type and denotes the smallest size of plant which is tracked by the model. Smaller plants are not considered, and their emergence from the recruitment processes is unresolved and therefore implicitly parameterized in the seedling establishment model.. The diameter of each cohort is then specified using the log-linear allometry between stem diameter and canopy height

$$dbh_{coh} = 10^{\frac{\log_{10}(h_{coh}) - c_{allom}}{m_{allom}}}$$

where the slope of the log-log relationship, m_{allom} is 0.64 and the intercept c_{allom} is 0.37. The structural biomass associated with a plant of this diameter and height is given (as a function of wood density, ρ , g cm⁻³)

$$b_{struc,coh} = c_{str} h_{coh}^{e_{str,hite}} dbh_{coh}^{e_{str,dbh}} \rho_{ft}^{e_{str,dens}}$$

taken from the original ED1.0 allometry [Moorecroft et al. 2001](#) (values of the allometric constants in Table *[table:allom]*). The maximum amount of leaf biomass associated with this diameter of tree is calculated according to the following allometry

$$b_{max,leaf,coh} = c_{leaf} dbh_{coh}^{e_{leaf,dbh}} \rho_{ft}^{e_{leaf,dens}}$$

from this quantity, we calculate the active/fine root biomass $b_{root,coh}$ as

$$b_{root,coh} = b_{max,leaf,coh} \cdot f_{frla}$$

where f_{frla} is the fraction of fine root biomass to leaf biomass, assigned per PFT

Table 2.41: Parameters needed for model initialization.

Parameter Symbol	Parameter Name	Units	Default Value
h_{min}	Minimum plant height	m	1.5
S_{init}	Initial Planting density	Individuals m ⁻²	
A_{tot}	Model area	m ²	10,000

[table:init]

2.28.4 Allocation of biomass

[6]_Total live biomass b_{alive} is the state variable of the model that describes the sum of the three live biomass pools leaf b_{leaf} , root b_{root} and sapwood b_{sw} (all in kGC individual⁻¹). The quantities are constrained by the following

$$b_{alive} = b_{leaf} + b_{root} + b_{sw}$$

Sapwood volume is a function of tree height and leaf biomass

$$b_{sw} = b_{leaf} \cdot h_{coh} \cdot f_{swh}$$

where f_{swh} is the ratio of sapwood mass (kgC) to leaf mass per unit tree height (m). Also, root mass is a function of leaf mass

$$b_{root} = b_{leaf} \cdot f_{swh}$$

Thus

$$b_{alive} = b_{leaf} + b_{leaf} \cdot f_{frla} + b_{leaf} \cdot h_{coh} \cdot f_{swh}$$

Rearranging gives the fraction of biomass in the leaf pool f_{leaf} as

$$f_{leaf} = \frac{1}{1 + h_{coh} \cdot f_{swh} + f_{frla}}$$

Thus, we can determine the leaf fraction from the height at the tissue ratios, and the phenological status of the cohort $S_{phen,coh}$.

$$b_{leaf} = b_{alive} \cdot l_{frac}$$

To divide the live biomass pool at restart, or whenever it is recalculated, into its constituent parts, we first

$$b_{leaf} = \begin{cases} b_{alive} \cdot l_{frac} & \text{for } S_{phen,coh} = 1 \\ 0 & \text{for } S_{phen,coh} = 0 \end{cases}$$

Because sometimes the leaves are dropped, using leaf biomass as a predictor of root and sapwood would produce zero live biomass in the winter. To account for this, we add the LAI memory variable l_{memory} to the live biomass pool to account for the need to maintain root biomass when leaf biomass is zero. Thus, to calculate the root biomass, we use

$$b_{root} = (b_{alive} + l_{memory}) \cdot l_{frac} \cdot f_{frla}$$

To calculate the sapwood biomass, we use

$$b_{sw} = (b_{alive} + l_{memory}) \cdot l_{frac} \cdot f_{swh} \cdot h_{coh}$$

Table 2.42: Allometric Constants

Parameter Symbol	Parameter Name	Units	Default Value
c_{allom}	Allometry intercept		0.37
m_{allom}	Allometry slope		0.64
c_{str}	Structural biomass multiplier		0.06896
$e_{str, dbh}$	Structural Biomass dbh exponent		1.94
$e_{str, hite}$	Structural Biomass height exponent		0.572
$e_{str, dens}$	Structural Biomass density exponent		0.931
c_{leaf}	Leaf biomass multiplier		0.0419
$e_{leaf, dbh}$	Leaf biomass dbh exponent		1.56
$e_{leaf, dens}$	Leaf biomass density exponent		0.55
f_{swh}	Ratio of sapwood mass to height	m^{-1}	
f_{frla}	Ratio of fine root mass to leaf mass	•	1.0

[table:allom]

2.28.5 Canopy Structure and the Perfect Plasticity Approximation

[7]_During initialization and every subsequent daily ED timestep, the canopy structure model is called to determine how the leaf area of the different cohorts is arranged relative to the incoming radiation, which will then be used to drive the radiation and photosynthesis calculations. This task requires that some assumptions are made about 1) the shape and depth of the canopy within which the plant leaves are arranged and 2) how the leaves of different cohorts are arranged relative to each other. This set of assumptions are critical to model performance in ED-like cohort based models, since they determine how light resources are partitioned between competing plants of varying heights, which has a very significant impact on how vegetation distribution emerges from competition *Fisher et al. 2010*.

The standard ED1.0 model makes a simple ‘flat disk’ assumption, that the leaf area of each cohort is spread in an homogenous layer at one exact height across entire the ground area represented by each patch. FATES has diverged from this representation due to (at least) two problematic emergent properties that we identified as generating unrealistic behaviours especially for large-area patches.

1. Over-estimation of light competition . The vertical stacking of cohorts which have all their leaf area at the same nominal height means that when one cohort is only very slightly taller than its competitor, it is completely shaded by it. This means that any small advantage in terms of height growth translates into a large advantage in terms of light competition, even at the seedling stage. This property of the model artificially exaggerates the process of light competition. In reality, trees do not compete for light until their canopies begin to overlap and canopy closure is approached.

2. Unrealistic over-crowding. The ‘flat-disk’ assumption has no consideration of the spatial extent of tree crowns. Therefore it has no control on the packing density of plants in the model. Given a mismatch between production and mortality, entirely unrealistic tree densities are thus possible for some combinations of recruitment, growth and mortality rates.

To account for the filling of space in three dimensions using the one-dimensional representation of the canopy employed by CLM, we implement a new scheme derived from that of *Purves et al. 2008*. Their argument follows the development of an individual-based variant of the SORTIE model, called SHELL, which allows the location of

individual plant crowns to be highly flexible in space. Ultimately, the solutions of this model possess an emergent property whereby the crowns of the plants simply fill all of the available space in the canopy before forming a distinct understorey.

Purves et al. developed a model that uses this feature, called the ‘perfect plasticity approximation’, which assumes the plants are able to perfectly fill all of the available canopy space. That is, at canopy closure, all of the available horizontal space is filled, with negligible gaps, owing to lateral tree growth and the ability of tree canopies to grow into the available gaps (this is of course, an over-simplified but potential useful ecosystem property). The ‘perfect plasticity approximation’ (PPA) implies that the community of trees is subdivided into discrete canopy layers, and by extension, each cohort represented by FATES model is assigned a canopy layer status flag, C_L . In this version, we set the maximum number of canopy layers at 2 for simplicity, although is possible to have a larger number of layers in theory. $C_{L,coh} = 1$ means that all the trees of cohort coh are in the upper canopy (overstory), and $C_{L,coh} = 2$ means that all the trees of cohort coh are in the understorey.

In this model, all the trees in the canopy experience full light on their uppermost leaf layer, and all trees in the understorey experience the same light (full sunlight attenuated by the average LAI of the upper canopy) on their uppermost leaves, as described in the radiation transfer section (more nuanced versions of this approach may be investigated in future model versions). The canopy is assumed to be cylindrical, the lower layers of which experience self-shading by the upper layers.

To determine whether a second canopy layer is required, the model needs to know the spatial extent of tree crowns. Crown area, A_{crown} , m², is defined as

$$A_{crown,coh} = \pi(dbh_{coh}S_{c,patch,Cl})^{1.56}$$

where A_{crown} is the crown area of a single tree canopy (m²) and $S_{c,patch,Cl}$ is the ‘canopy spread’ parameter (m cm⁻¹) of this canopy layer, which is assigned as a function of canopy space filling, discussed below. In contrast to [Purves et al. 2008](#), we use an exponent, identical to that for leaf biomass, of 1.56, not 2.0, such that tree leaf area index does not change as a function of diameter.

To determine whether the canopy is closed, we calculate the total canopy area as:

$$A_{canopy} = \sum_{coh=1}^{nc,patch} A_{crown,coh} \cdot n_{coh}$$

where nc_{patch} is the number of cohorts in a given patch. If the area of all crowns A_{canopy} (m²) is larger than the total ground area of a patch (A_{patch}), then some fraction of each cohort is demoted to the understorey.

Under these circumstances, the *extra* crown area A_{loss} (i.e., $A_{canopy} - A_p$) is moved into the understorey. For each cohort already in the canopy, we determine a fraction of trees that are moved from the canopy (L_c) to the understorey. L_c is calculated as [Fisher et al. 2010](#)

$$L_c = \frac{A_{loss,patch}w_{coh}}{\sum_{coh=1}^{nc,patch} w_{coh}},$$

where w_{coh} is a weighting of each cohort determined by basal diameter dbh (cm) and the competitive exclusion coefficient C_e

$$w_{coh} = dbh_{coh}C_e.$$

The higher the value of C_e the greater the impact of tree diameter on the probability of a given tree obtaining a position in the canopy layer. That is, for high C_e values, competition is highly deterministic. The smaller the value of C_e , the greater the influence of random factors on the competitive exclusion process, and the higher the probability that slower growing trees will get into the canopy. Appropriate values of C_e are poorly constrained but alter the outcome of competitive processes.

The process by which trees are moved between canopy layers is complex because 1) the crown area predicted for a cohort to lose may be larger than the total crown area of the cohort, which requires iterative solutions, and 2) on some

occasions (e.g. after fire), the canopy may open up and require ‘promotion’ of cohorts from the understorey, and 3) canopy area may change due to the variations of canopy spread values ($S_{c,patch,Cl}$, see the section below for details) when fractions of cohorts are demoted or promoted. Further details can be found in the code references in the footnote.

Horizontal Canopy Spread

⁸Purves et al. 2008 estimated the ratio between canopy and stem diameter c_p as 0.1 m cm^{-1} for canopy trees in North American forests, but this estimate was made on trees in closed canopies, whose shape is subject to space competition from other individuals. Sapling trees have no constraints in their horizontal spatial structure, and as such, are more likely to display their leaves to full sunlight. Also, prior to canopy closure, light interception by leaves on the sides of the canopy is also higher than it would be in a closed canopy forest. If the ‘canopy spread’ parameter is constant for all trees, then we simulate high levels of self-shading for plants in unclosed canopies, which is arguably unrealistic and can lower the productivity of trees in areas of unclosed canopy (e.g. low productivity areas of boreal or semi-arid regions where LAI and canopy cover might naturally be low). We here interpret the degree of canopy spread, S_c as a function of how much tree crowns interfere with each other in space, or the total canopy area A_{canopy} . However A_{canopy} itself is a function of S_c , leading to a circularity. S_c is thus solved iteratively through time.

Each daily model step, A_{canopy} and the fraction of the gridcell occupied by tree canopies in each canopy layer ($A_{f,Cl} = A_{canopy,Cl}/A_{patch}$) is calculated based on S_c from the previous timestep. If A_f is greater than a threshold value A_t , S_c is increased by a small increment i . The threshold A_t is, hypothetically, the canopy fraction at which light competition begins to impact on tree growth. This is less than 1.0 owing to the non-perfect spatial spacing of tree canopies. If $A_{f,Cl}$ is greater than A_t , then S_c is reduced by an increment i , to reduce the spatial extent of the canopy, thus.

$$S_{c,patch,Cl,t+1} = \begin{cases} S_{c,patch,Cl,t} + i & \text{for } A_{f,Cl} < A_t \\ S_{c,patch,Cl,t} - i & \text{for } A_{f,Cl} > A_t \end{cases}$$

The values of S_c are bounded to upper and lower limits. The lower limit corresponds to the observed canopy spread parameter for canopy trees $S_{c,min}$ and the upper limit corresponds to the largest canopy extent $S_{c,max}$

$$S_{c,patch,Cl} = \begin{cases} S_{c,min} & \text{for } S_{c,patch,Cl} < S_{c,min} \\ S_{c,max} & \text{for } S_{c,patch,Cl} > S_{c,max} \end{cases}$$

This iterative scheme requires two additional parameters (i and A_t). i affects the speed with which canopy spread (and hence leaf area index) increase as canopy closure is neared. However, the model is relatively insensitive to the choice of either i or A_t .

Definition of Leaf and Stem Area Profile

[9]_Within each patch, the model defines and tracks cohorts of multiple plant functional types that exist either in the canopy or understorey. Light on the top leaf surface of each cohort in the canopy is the same, and the rate of decay through the canopy is also the same for each PFT. Therefore, we accumulate all the cohorts of a given PFT together for the sake of the radiation and photosynthesis calculations (to avoid separate calculations for every cohort).

Therefore, the leaf area index for each patch is defined as a three-dimensional array $lai_{Cl,ft,z}$ where Cl is the canopy layer, ft is the functional type and z is the leaf layer within each canopy. This three-dimensional structure is the basis of the radiation and photosynthetic models. In addition to a leaf area profile matrix, we also define, for each patch, the area which is covered by leaves at each layer as $carea_{Cl,ft,z}$.

Each plant cohort is already defined as a member of a single canopy layer and functional type. This means that to generate the $x_{Cl,ft,z}$ matrix, it only remains to divide the leaf area of each cohort into leaf layers. First, we determine

⁸ This description relates to algorithms in the canopy_spread subroutine

how many leaf layers are occupied by a single cohort, by calculating the ‘tree LAI’ as the total leaf area of each cohort divided by its crown area (both in m²)

$$tree_{lai,coh} = \frac{b_{leaf,coh} \cdot sla_{ft}}{A_{crown,coh}}$$

where sla_{ft} is the specific leaf area in m² KgC⁻¹ and b_{leaf} is in kGC per plant.

Stem area index (SAI) is ratio of the total area of all woody stems on a plant to the area of ground covered by the plant. During winter in deciduous areas, the extra absorption by woody stems can have a significant impact on the surface energy budget. However, in previous *big leaf* versions of the CLM, computing the circumstances under which stem area was visible in the absence of leaves was difficult and the algorithm was largely heuristic as a result. Given the multi-layer canopy introduced for FATES, we can determine the leaves in the higher canopy layers will likely shade stem area in the lower layers when leaves are on, and therefore stem area index can be calculated as a function of woody biomass directly.

Literature on stem area index is particularly poor, as it’s estimation is complex and not particularly amenable to the use of, for example, assumptions of random distribution in space that are typically used to calculate leaf area from light interception. *Kucharik et al. 1998* estimated that SAI visible from an LAI2000 sensor was around 0.5 m² m⁻². Low et al. 2001 estimate that the wood area index for Ponderosa Pine forest is 0.27-0.33. The existing CLM(CN) algorithm sets the minimum SAI at 0.25 to match MODIS observations, but then allows SAI to rise as a function of the LAI lost, meaning than in some places, predicted SAI can reach value of 8 or more. Clearly, greater scientific input on this quantity is badly needed. Here we determine that SAI is a linear function of woody biomass, to at very least provide a mechanistic link between the existence of wood and radiation absorbed by it. The non-linearity between how much woody area exists and how much radiation is absorbed is provided by the radiation absorption algorithm. Specifically, the SAI of an individual cohort ($tree_{sai,coh}$, m² m⁻²) is calculated as follows,

$$tree_{sai,coh} = k_{sai} \cdot b_{struc,coh},$$

where k_{sai} is the coefficient linking structural biomass to SAI. The number of occupied leaf layers for cohort coh ($n_{z,coh}$) is then equal to the rounded up integer value of the tree SAI ($tree_{sai,coh}$) and LAI ($tree_{lai,coh}$) divided by the layer thickness (i.e., the resolution of the canopy layer model, in units of vegetation index ($lai+sai$) with a default value of 1.0, δ_{vai}),

$$n_{z,coh} = \frac{tree_{lai,coh} + tree_{sai,coh}}{\delta_{vai}}.$$

The fraction of each layer that is leaf (as opposed to stem) can then be calculated as

$$f_{leaf,coh} = \frac{tree_{lai,coh}}{tree_{sai,coh} + tree_{lai,coh}}.$$

Finally, the leaf area in each leaf layer pertaining to this cohort is thus

$$lai_{z,coh} = \begin{cases} \delta_{vai} \cdot f_{leaf,coh} \frac{A_{canopy,coh}}{A_{canopy,patch}} & \text{for } i = 1, \dots, i = n_{z,coh} - 1 \\ \delta_{vai} \cdot f_{leaf,coh} \frac{A_{canopy,coh}}{A_{canopy,patch}} \cdot r_{vai} & \text{for } i = n_{z,coh} \end{cases}$$

and the stem area index is

$$sai_{z,coh} = \begin{cases} \delta_{vai} \cdot (1 - f_{leaf,coh}) \frac{A_{canopy,coh}}{A_{canopy,patch}} & \text{for } i = 1, \dots, i = n_{z,coh} - 1 \\ \delta_{vai} \cdot (1 - f_{leaf,coh}) \frac{A_{canopy,coh}}{A_{canopy,patch}} \cdot r_{vai} & \text{for } i = n_{z,coh} \end{cases}$$

where r_{vai} is the remainder of the canopy that is below the last full leaf layer

$$r_{vai} = (tree_{lai,coh} + tree_{sai,coh}) - (\delta_{vai} \cdot (n_{z,coh} - 1)).$$

$A_{canopy,patch}$ is the total canopy area occupied by plants in a given patch (m²) and is calculated as follows,

$$A_{canopy,patch} = \min \left(\sum_{coh=1}^{coh=ncoh} A_{canopy,coh}, A_{patch} \right).$$

The canopy is conceived as a cylinder, although this assumption could be altered given sufficient evidence that canopy shape was an important determinant of competitive outcomes, and the area of ground covered by each leaf layer is the same through the cohort canopy. With the calculated SAI and LAI, we are able to calculate the complete canopy profile. Specifically, the relative canopy area for the cohort coh is calculated as

$$area_{1:nz,coh} = \frac{A_{crown,coh}}{A_{canopy,patch}}.$$

The total occupied canopy area for each canopy layer (Cl), plant functional type (ft) and leaf layer (z) bin is thus

$$carea_{Cl,ft,z} = \sum_{coh=1}^{coh=ncoh} area_{1:nz,coh}$$

where $ft_{coh} = ft$ and $Cl_{coh} = Cl$.

All of these quantities are summed across cohorts to give the complete leaf and stem area profiles,

$$lai_{Cl,ft,z} = \sum_{coh=1}^{coh=ncoh} lai_{z,coh}$$

$$sai_{Cl,ft,z} = \sum_{coh=1}^{coh=ncoh} sai_{z,coh}$$

Burial of leaf area by snow

The calculations above all pertain to the total leaf and stem area indices which characterize the vegetation structure. In addition, the model must know when the vegetation is covered by snow, and by how much, so that the albedo and energy balance calculations can be adjusted accordingly. Therefore, we calculated a ‘total’ and ‘exposed’ lai and sai profile using a representation of the bottom and top canopy heights, and the depth of the average snow pack. For each leaf layer z of each cohort, we calculate an ‘exposed fraction $f_{exp,z}$ ’ via consideration of the top and bottom heights of that layer $h_{top,z}$ and $h_{bot,z}$ (m),

$$h_{top,z} = h_{coh} - h_{coh} \cdot f_{crown,ft} \cdot \frac{z}{n_{z,coh}}$$

$$h_{bot,z} = h_{coh} - h_{coh} \cdot f_{crown,ft} \cdot \frac{z+1}{n_{z,coh}}$$

where $f_{crown,ft}$ is the plant functional type (ft) specific fraction of the cohort height that is occupied by the crown. Specifically, the ‘exposed fraction $f_{exp,z}$ ’ is calculated as follows,

$$f_{exp,z} = \begin{cases} 1.0 & h_{bot,z} > d_{snow} \\ \frac{d_{snow} - h_{bot,z}}{h_{top,z} - h_{bot,z}} & h_{top,z} > d_{snow}, h_{bot,z} < d_{snow} \\ 0.0 & h_{top,z} < d_{snow} \end{cases}$$

The resulting exposed ($elai$, $esai$) and total ($tlai$, $tsai$) leaf and stem area indices are calculated as

$$\begin{aligned} elai_{Cl,ft,z} &= lai_{Cl,ft,z} \cdot f_{exp,z} \\ esai_{Cl,ft,z} &= sai_{Cl,ft,z} \cdot f_{exp,z} \\ tlai_{Cl,ft,z} &= lai_{Cl,ft,z} \\ tsai_{Cl,ft,z} &= sai_{Cl,ft,z} \end{aligned}$$

and are used in the radiation interception and photosynthesis algorithms described later.

Parameter Symbol	Parameter Name	Units	Notes	Indexed by
δ_{vai}	Thickness of single canopy layer	$m^{-2} m: math.^{-2}$		
C_e	Competitive Exclusion Parameter	none		
$c_{p,min}$	Minimum canopy spread	$m^2 cm: math.^{-1}$		
$c_{p,max}$	Competitive Exclusion Parameter	$m^2 cm: math.^{-1}$		
i	Incremental change in c_p	$m^2 cm: math.^{-1} y^{-1}$		
A_t	Threshold canopy closure	none		
$f_{crown,ft}$	Crown fraction	none		ft
k_{sai}	Stem area per unit woody biomass	$m^2 KgC^{-1}$		

2.28.6 Radiation Transfer

Fundamental Radiation Transfer Theory

[10]_The first interaction of the land surface with the properties of vegetation concerns the partitioning of energy into that which is absorbed by vegetation, reflected back into the atmosphere, and absorbed by the ground surface. Older versions of the CLM have utilized a “two-stream” approximation [Sellers 1985](#), [Sellers et al. 1986](#) that provided an empirical solution for the radiation partitioning of a multi-layer canopy for two streams, of diffuse and direct light. However, implementation of the Ecosystem Demography model requires a) the adoption of an explicit multiple layer canopy b) the implementation of a multiple plant type canopy and c) the distinction of canopy and under-storey layers, in-between which the radiation streams are fully mixed. The radiation mixing between canopy layers is necessary as the position of different plants in the under-storey is not defined spatially or relative to the canopy trees above. In this new scheme, we thus implemented a one-dimensional scheme that traces the absorption, transmittance and reflectance of each canopy layer and the soil, iterating the upwards and downwards passes of radiation through the canopy until a pre-defined accuracy tolerance is reached. This approach is based on the work of [Norman 1979](#).

Here we describe the basic theory of the radiation transfer model for the case of a single homogenous canopy, and in the next section we discuss how this is applied to the multi layer multi PFT canopy in the FATES implementation. The code considers the fractions of a single unit of incoming direct and a single unit of incoming diffuse light, that are absorbed at each layer of the canopy for a given solar angle (α_s , radians). Direct radiation is extinguished through the canopy according to the coefficient k_{dir} that is calculated from the incoming solar angle and the dimensionless leaf angle distribution parameter (χ) as

$$k_{dir} = g_{dir} / \sin(\alpha_s)$$

where

$$g_{dir} = \phi_1 + \phi_2 \cdot \sin(\alpha_s)$$

and

$$\begin{aligned}\phi_1 &= 0.5 - 0.633\chi_l - 0.33\chi_l^2 \\ \phi_2 &= 0.877(1 - 2\phi_1)\end{aligned}$$

The leaf angle distribution is a descriptor of how leaf surfaces are arranged in space. Values approaching 1.0 indicate that (on average) the majority of leaves are horizontally arranged with respect to the ground. Values approaching -1.0 indicate that leaves are mostly vertically arranged, and a value of 0.0 denotes a canopy where leaf angle is random (a ‘spherical’ distribution).

According to Beer’s Law, the fraction of light that is transferred through a single layer of vegetation (leaves or stems) of thickness δ_{vai} , without being intercepted by any surface, is

$$tr_{dir} = e^{-k_{dir}\delta_{vai}}$$

and the incident direct radiation transmitted to each layer of the canopy ($dir_{tr,z}$) is thus calculated from the cumulative leaf area (L_{above}) shading each layer (z):

$$dir_{tr,z} = e^{-k_{dir}L_{above,z}}$$

The fraction of the leaves f_{sun} that are exposed to direct light is also calculated from the decay coefficient k_{dir} .

$$f_{sun,z} = e^{-k_{dir}L_{above,z}}$$

and

$$f_{shade,z} = 1 - f_{sun,z}$$

where $f_{shade,z}$ is the fraction of leaves that are shaded from direct radiation and only receive diffuse light.

Diffuse radiation, by definition, enters the canopy from a spectrum of potential incident directions, therefore the un-intercepted transfer (tr_{dif}) through a leaf layer of thickness δ_l is calculated as the mean of the transfer rate from each of 9 different incident light directions (α_s) between 0 and 180 degrees to the horizontal.

$$tr_{dif} = \frac{1}{9} \sum_{\alpha_s=5\pi/180}^{\alpha_s=85\pi/180} e^{-k_{dir,l}\delta_{vai}}$$

$$tr_{dif} = \frac{1}{9}\pi \sum_{\alpha_s=0}^{\pi/2} \frac{e^{-gdir}\alpha_s}{\delta_{vai} \cdot \sin(\alpha_s)\sin(\alpha_s)\cos(\alpha_s)}$$

The fraction $(1-tr_{dif})$ of the diffuse radiation is intercepted by leaves as it passes through each leaf layer. Of this, some fraction is reflected by the leaf surfaces and some is transmitted through. The fractions of diffuse radiation reflected from ($refl_{dif}$) and transmitted though ($tran_{dif}$) each layer of leaves are thus, respectively

$$\begin{aligned} refl_{dif} &= (1 - tr_{dif})\rho_{l,ft} \\ tran_{dif} &= (1 - tr_{dif})\tau_{l,ft} + tr_{dif} \end{aligned}$$

where $\rho_{l,ft}$ and $\tau_{l,ft}$ are the fractions of incident light reflected and transmitted by individual leaf surfaces.

Once we know the fractions of light that are transmitted and reflected by each leaf layer, we begin the process of distributing light through the canopy. Starting with the first leaf layer ($z=1$), where the incident downwards diffuse radiation (dif_{down}) is 1.0, we work downwards for n_z layers, calculating the radiation in the next layer down ($z+1$) as:

$$dif_{down,z+1} = \frac{dif_{down,z}tran_{dif}}{1 - r_{z+1}refl_{dif}}$$

Here, $dif_{down,z}tran_{dif}$ calculates the fraction of incoming energy transmitted downwards onto layer $z+1$. This flux is then increased by the additional radiation r_z that is reflected upwards from further down in the canopy to layer z , and then is reflected back downwards according to the reflected fraction $refl_{dif}$. The more radiation in $r_{z+1}refl_{dif}$, the smaller the denominator and the larger the downwards flux. r is also calculated sequentially, starting this time at the soil surface layer (where $z = n_z + 1$)

$$r_{n_z+1} = alb_s$$

where alb_s is the soil albedo characteristic. The upwards reflected fraction r_z for each leaf layer, moving upwards, is then [Norman 1979](#)

$$r_z = \frac{r_{z+1} \times tran_{dif}^2}{(1 - r_{z+1}refl_{dif}) + refl_{dif}}.$$

The corresponding upwards diffuse radiation flux is therefore the fraction of downwards radiation that is incident on a particular layer, multiplied by the fraction that is reflected from all the lower layers:

$$dif_{up,z} = r_z dif_{down,z+1}$$

Now we have initial conditions for the upwards and downwards diffuse fluxes, these must be modified to account for the fact that, on interception with leaves, direct radiation is transformed into diffuse radiation. In addition, the initial solutions to the upwards and downwards radiation only allow a single ‘bounce’ of radiation through the canopy, so some radiation which might be intercepted by leaves higher up is potentially lost. Therefore, the solution to this model is iterative. The iterative solution has upwards and a downwards components that calculate the upwards and downwards fluxes of total radiation at each leaf layer ($rad_{dn,z}$ and $rad_{up,z}$). The downwards component begins at the top canopy layer ($z = 1$). Here we define the incoming solar diffuse and direct radiation ($solar_{dir}$ and $solar_{dif}$ respectively).

$$dif_{dn,1} = solar_{dif}$$

$$rad_{dn,z+1} = dif_{dn,z} \cdot trans_{dif} + dif_{up,z+1} \cdot refl_{dif} + solar_{dir} \cdot dir_{tr,z} (1 - tr_{dir}) \tau_l.$$

The first term of the right-hand side deals with the diffuse radiation transmitted downwards, the second with the diffuse radiation travelling upwards, and the third with the direct radiation incoming at each layer ($dir_{tr,z}$) that is intercepted by leaves ($1 - tr_{dir}$) and then transmitted through through the leaf matrix as diffuse radiation (τ_l). At the bottom of the canopy, the light reflected off the soil surface is calculated as

$$rad_{up,nz} = dif_{down,z} \cdot salb_{dif} + solar_{dir} \cdot dir_{tr,z} salb_{dir}.$$

The upwards propagation of the reflected radiation is then

$$rad_{up,z} = dif_{up,z+1} \cdot trans_{dif} + dif_{dn,z} \cdot refl_{dif} + solar_{dir} \cdot dir_{tr,z} (1 - tr_{dir}) \rho_l.$$

Here the first two terms deal with the diffuse downwards and upwards fluxes, as before, and the third deals direct beam light that is intercepted by leaves and reflected upwards. These upwards and downwards fluxes are computed for multiple iterations, and at each iteration, $rad_{up,z}$ and $rad_{down,z}$ are compared to their values in the previous iteration. The iteration scheme stops once the differences between iterations for all layers is below a predefined tolerance factor, (set here at 10^{-4}). Subsequently, the fractions of absorbed direct ($abs_{dir,z}$) and diffuse ($abs_{dif,z}$) radiation for each leaf layer then

$$abs_{dir,z} = solar_{dir} \cdot dir_{tr,z} \cdot (1 - tr_{dir}) \cdot (1 - \rho_l - \tau_l)$$

$$abs_{dif,z} = (dif_{dn,z} + dif_{up,z+1}) \cdot (1 - tr_{dif}) \cdot (1 - \rho_l - \tau_l).$$

and, the radiation energy absorbed by the soil for the diffuse and direct streams is is calculated as

$$abs_{soil} = dif_{down,nz+1} \cdot (1 - salb_{dif}) + solar_{dir} \cdot dir_{tr,nz+1} \cdot (1 - salb_{dir}).$$

Canopy level albedo is denoted as the upwards flux from the top leaf layer

$$alb_{canopy} = \frac{dif_{up,z+1}}{solar_{dir} + solar_{dif}}$$

and the division of absorbed energy into sunlit and shaded leaf fractions, (required by the photosynthesis calculations), is

$$abs_{sha,z} = abs_{dif,z} \cdot f_{sha}$$

$$abs_{sun,z} = abs_{dif,z} \cdot f_{sun} + abs_{dir,z}$$

Resolution of radiation transfer theory within the FATES canopy structure

The radiation transfer theory above, was described with reference to a single canopy of one plant functional type, for the sake of clarity of explanation. The FATES model, however, calculates radiative and photosynthetic fluxes for a more complex hierarchical structure within each patch/time-since-disturbance class, as described in the leaf area

profile section. Firstly, we denote two or more canopy layers (denoted C_l). The concept of a ‘canopy layer’ refers to the idea that plants are organized into discrete over and under-stories, as predicted by the Perfect Plasticity Approximation ([Purves et al. 2008](#), [Fisher et al. 2010](#)). Within each canopy layer there potentially exist multiple cohorts of different plant functional types and heights. Within each canopy layer, C_l , and functional type, ft , the model resolves numerous leaf layers z , and, for some processes, notably photosynthesis, each leaf layer is split into a fraction of sun and shade leaves, f_{sun} and f_{sha} , respectively.

The radiation scheme described in Section is solved explicitly for this structure, for both the visible and near-infrared wavebands, according to the following assumptions.

- A *canopy layer* (C_L) refers to either the over or understorey
- A *leaf layer* (z) refers to the discretization of the LAI within the canopy of a given plant functional type.
- All PFTs in the same canopy layer have the same solar radiation incident on the top layer of the canopy
- Light is transmitted through the canopy of each plant functional type independently
- Between canopy layers, the light streams from different plant functional types are mixed, such that the (undefined) spatial location of plants in lower canopy layers does not impact the amount of light received.
- Where understorey layers fill less area than the overstorey layers, radiation is directly transferred to the soil surface.
- All these calculations pertain to a single patch, so we omit the *patch* subscript for simplicity in the following discussion.

Within this framework, the majority of the terms in the radiative transfer scheme are calculated with indices of C_L , ft and z . In the following text, we revisit the simplified version of the radiation model described above, and explain how it is modified to account for the more complex canopy structure used by FATES.

Firstly, the light penetration functions, k_{dir} and g_{dir} are described as functions of ft , because the leaf angle distribution, χ_l , is a pft-specific parameter. Thus, the diffuse irradiance transfer rate, tr_{dif} is also ft specific because g_{dir} , on which it depends, is a function of χ_l .

The amount of direct light reaching each leaf layer is a function of the leaves existing above the layer in question. If a leaf layer ‘ z ’ is in the top canopy layer (the over-storey), it is only shaded by leaves of the same PFT so k_{dir} is unchanged from equation. If there is more than one canopy layer ($C_{l,max} > 1$), then the amount of direct light reaching the top leaf surfaces of the second/lower layer is the weighted average of the light attenuated by all the parallel tree canopies in the canopy layer above, thus.

$$dir_{tr,C_l,:,:1} = \sum_{ft=1}^{npft} (dir_{tr,C_l,ft,z_{max}} \cdot c_{area,C_l-1,ft,z_{max}})$$

where pft_{wt} is the areal fraction of each canopy layer occupied by each functional type and z_{max} is the index of the bottom canopy layer of each pft in each canopy layer (the subscripts C_l and ft are implied but omitted from all z_{max} references to avoid additional complications)

Similarly, the sunlit fraction for a leaf layer ‘ z ’ in the second canopy layer (where $C_l > 1$) is

$$f_{sun,C_l,ft,z} = W_{sun,C_l} \cdot e^{k_{dir,ft,laic,z}}$$

where W_{sun,C_l} is the weighted average sunlit fraction in the bottom layer of a given canopy layer.

$$W_{sun,C_l} = \sum_{ft=1}^{npft} (f_{sun,C_l-1,ft,z_{max}} \cdot c_{area,C_l-1,ft,z_{max}})$$

Following through the sequence of equations for the simple single pft and canopy layer approach above, the $refl_{dif}$ and $tran_{dif}$ fluxes are also indexed by C_l , ft , and z . The diffuse radiation reflectance ratio r_z is also calculated in a manner that homogenizes fluxes between canopy layers. For the canopy layer nearest the soil ($C_l = C_{l,max}$). For the

top canopy layer ($C_l=1$), a weighted average reflectance from the lower layers is used as the baseline, in lieu of the soil albedo. Thus:

$$r_{z,Cl,:,1} = \sum_{ft=1}^{npft} (r_{z,Cl-1,ft,1} pft_{wt,Cl-1,ft,1})$$

For the iterative flux resolution, the upwards and downwards fluxes are also averaged between canopy layers, thus where $C_l > 1$

$$rad_{dn,Cl,ft,1} = \sum_{ft=1}^{npft} (rad_{dn,Cl-1,ft,zmax} \cdot pft_{wt,Cl-1,ft,zmax})$$

and where $C_l = 1$, and $C_{l,max} > 1$

$$rad_{up,Cl,ft,zmax} = \sum_{ft=1}^{npft} (rad_{up,Cl+1,ft,1} \cdot pft_{wt,Cl+1,ft,1})$$

The remaining terms in the radiation calculations are all also indexed by C_l , ft and z so that the fraction of absorbed radiation outputs are termed $abs_{dir,Cl,ft,z}$ and $abs_{dif,Cl,ft,z}$. The sunlit and shaded absorption rates are therefore

$$abs_{sha,Cl,ft,z} = abs_{dif,Cl,ft,z} \cdot f_{sha,Cl,ft,z}$$

and

$$abs_{sun,Cl,ft,z} = abs_{dif,Cl,ft,z} \cdot f_{sun,Cl,ft,z} + abs_{dir,Cl,ft,z}$$

The albedo of the mixed pft canopy is calculated as the weighted average of the upwards radiation from the top leaf layer of each pft where $C_l=1$:

$$alb_{canopy} = \sum_{ft=1}^{npft} \frac{dif_{up,1,ft,1} pft_{wt,1,ft,1}}{solar_{dir} + solar_{dif}}$$

The radiation absorbed by the soil after passing through under-storey vegetation is:

$$abs_{soil} = \sum_{ft=1}^{npft} pft_{wt,1,ft,1} (dif_{down,nz+1} (1 - salb_{dif}) + solar_{dir} dir_{tr,nz+1} (1 - salb_{dir}))$$

to which is added the diffuse flux coming directly from the upper canopy and hitting no understorey vegetation.

$$abs_{soil} = abs_{soil} + dif_{dn,2,1} (1 - \sum_{ft=1}^{npft} pft_{wt,1,ft,1}) (1 - salb_{dif})$$

and the direct flux coming directly from the upper canopy and hitting no understorey vegetation.

$$abs_{soil} = abs_{soil} + solar_{dir} dir_{tr,2,1} (1 - \sum_{ft=1}^{npft} pft_{wt,1,ft,1}) (1 - salb_{dir})$$

These changes to the radiation code are designed to be structurally flexible, and the scheme may be collapsed down to only include one canopy layer, functional type and pft for testing if necessary.

Table 2.43: Parameters needed for radiation transfer model.

Parameter Symbol	Parameter Name	Units	indexed by
χ	Leaf angle distribution parameter	none	ft
ρ_l	Fraction of light reflected by leaf surface	none	ft
τ_l	Fraction of light transmitted by leaf surface	none	ft
alb_s	Fraction of light reflected by soil	none	direct vs diffuse

2.28.7 Photosynthesis

Fundamental photosynthetic physiology theory

[11] In this section we describe the physiological basis of the photosynthesis model before describing its application to the FATES canopy structure. This description in this section is largely repeated from the Oleson et al. CLM4.5 technical note but included here for comparison with its implementation in FATES. Photosynthesis in C3 plants is based on the model of [Farquhar 1980](#) as modified by [Collatz et al. 1991](#). Photosynthetic assimilation in C4 plants is based on the model of [Collatz et al. 1991](#). In both models, leaf photosynthesis, gpp ($\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$) is calculated as the minimum of three potentially limiting fluxes, described below:

$$\text{gpp} = \min(w_j, w_c, w_p).$$

The RuBP carboxylase (Rubisco) limited rate of carboxylation w_c ($\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$) is determined as

$$w_c = \begin{cases} \frac{V_{c,max}(c_i - \Gamma_*)}{c_i + K_c(1 + o_i/K_o)} & \text{for C}_3 \text{ plants} \\ V_{c,max} & \text{for C}_4 \text{ plants} \end{cases} \quad c_i - \Gamma_* \geq 0$$

where c_i is the internal leaf CO_2 partial pressure (Pa) and $o_i(0.209P_{atm})$ is the O_2 partial pressure (Pa). K_c and K_o are the Michaelis-Menten constants (Pa) for CO_2 and O_2 . These vary with vegetation temperature T_v ($^{\circ}\text{C}$) according to an Arrhenius function described in [Oleson et al. 2013](#). $V_{c,max}$ is the leaf layer photosynthetic capacity ($\mu\text{ mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$).

The maximum rate of carboxylation allowed by the capacity to regenerate RuBP (i.e., the light-limited rate) w_j ($\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$) is

$$w_j = \begin{cases} \frac{J(c_i - \Gamma_*)}{4c_i + 8\Gamma_*} & \text{for C}_3 \text{ plants} \\ 4.6\phi\alpha & \text{for C}_4 \text{ plants} \end{cases} \quad c_i - \Gamma_* \geq 0$$

To find J , the electron transport rate ($\mu\text{ mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$), we solve the following quadratic term and take its smaller root,

$$\Theta_{psII} J^2 - (I_{psII} + J_{max})J + I_{psII} J_{max} = 0$$

where J_{max} is the maximum potential rate of electron transport ($\mu\text{mol m}^{-2} \text{ s}^{-1}$), I_{PSII} is the light utilized in electron transport by photosystem II ($\mu\text{mol m}^{-2} \text{ s}^{-1}$) and Θ_{PSII} is curvature parameter. I_{PSII} is determined as

$$I_{PSII} = 0.5\Phi_{PSII}(4.6\phi)$$

where ϕ is the absorbed photosynthetically active radiation (W m^{-2}) for either sunlit or shaded leaves (abs_{sun} and abs_{sha}). ϕ is converted to photosynthetic photon flux assuming 4.6 μmol photons per joule. Parameter values are $\Phi_{PSII} = 0.7$ for C3 and $\Phi_{PSII} = 0.85$ for C4 plants.

The export limited rate of carboxylation for C3 plants and the PEP carboxylase limited rate of carboxylation for C4 plants w_e (also in $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$) is

$$w_e = \begin{cases} 3T_{p,0} & \text{for C}_3 \text{ plants} \\ k_p \frac{c_i}{P_{atm}} & \text{for C}_4 \text{ plants.} \end{cases}$$

T_p is the triose-phosphate limited rate of photosynthesis, which is equal to $0.167V_{c,max}0$. k_p is the initial slope of C4 CO_2 response curve. The Michaelis-Menten constants K_c and K_o are modeled as follows,

$$K_c = K_{c,25}(a_{kc})^{\frac{T_v - 25}{10}},$$

$$K_o = K_{o,25} (a_{ko})^{\frac{T_v - 25}{10}},$$

where $K_{c,25} = 30.0$ and $K_{o,25} = 30000.0$ are values (Pa) at 25 °C, and $a_{kc} = 2.1$ and $a_{ko} = 1.2$ are the relative changes in $K_{c,25}$ and $K_{o,25}$ respectively, for a 10°C change in temperature. The CO₂ compensation point Γ_* (Pa) is

$$\Gamma_* = \frac{1}{2} \frac{K_c}{K_o} 0.21 o_i$$

where the term 0.21 represents the ratio of maximum rates of oxygenation to carboxylation, which is virtually constant with temperature *Farquhar, 1980*.

Resolution of the photosynthesis theory within the FATES canopy structure.

The photosynthesis scheme is modified from the CLM4.5 model to give estimates of photosynthesis, respiration and stomatal conductance for a three dimensional matrix indexed by canopy level (C_l), plant functional type (ft) and leaf layer (z). We conduct the photosynthesis calculations at each layer for both sunlit and shaded leaves. Thus, the model also generates estimates of w_c , w_j and w_e indexed in the same three dimensional matrix. In this implementation, some properties (stomatal conductance parameters, top-of-canopy photosynthetic capacity) vary with plant functional type, and some vary with both functional type and canopy depth (absorbed photosynthetically active radiation, nitrogen-based variation in photosynthetic properties). The remaining drivers of photosynthesis (P_{atm} , K_c , o_i , K_o , temperature, atmospheric CO₂) remain the same throughout the canopy. The rate of gross photosynthesis ($gpp_{Cl,ft,z}$) is the smoothed minimum of the three potentially limiting processes (carboxylation, electron transport, export limitation), but calculated independently for each leaf layer:

$$gpp_{Cl,ft,z} = \min(w_{c,Cl,ft,z}, w_{j,Cl,ft,z}, w_{e,Cl,ft,z}).$$

For $w_{c,Cl,ft,z}$, we use

$$w_{c,Cl,ft,z} = \begin{cases} \frac{V_{c,max,Cl,ft,z}(c_{i,Cl,ft,z} - \Gamma_*)}{c_{i,Cl,ft,z} + K_c(1+o_i/K_o)} & \text{for } C_3 \text{ plants} \\ V_{c,max,Cl,ft,z} & c_{i,Cl,ft,z} - \Gamma_* \geq 0 \\ & \text{for } C_4 \text{ plants} \end{cases}$$

where $V_{c,max}$ now varies with PFT, canopy depth and layer (see below). Internal leaf CO₂ ($c_{i,Cl,ft,z}$) is tracked separately for each leaf layer. For the light limited rate w_j , we use

$$w_j = \begin{cases} \frac{J(c_i - \Gamma_*) 4.6 \phi \alpha}{4c_i + 8\Gamma_*} & \text{for } C_3 \text{ plants} \\ 4.6 \phi \alpha & \text{for } C_4 \text{ plants} \end{cases}$$

where J is calculated as above but based on the absorbed photosynthetically active radiation ($\phi_{Cl,ft,z}$) for either sunlit or shaded leaves in W m⁻². Specifically,

$$\phi_{Cl,ft,z} = \begin{cases} abs_{sun,Cl,ft,z} & \text{for sunlit leaves} \\ abs_{sha,Cl,ft,z} & \text{for shaded leaves} \end{cases}$$

The export limited rate of carboxylation for C3 plants and the PEP carboxylase limited rate of carboxylation for C4 plants w_e (also in μmol CO₂ m⁻² s⁻¹) is calculated in a similar fashion,

$$w_{e,Cl,ft,z} = \begin{cases} 0.5V_{c,max,Cl,ft,z} & \text{for } C_3 \text{ plants} \\ 4000V_{c,max,Cl,ft,z} \frac{c_{i,Cl,ft,z}}{P_{atm}} & \text{for } C_4 \text{ plants.} \end{cases}$$

Variation in plant physiology with canopy depth

Both $V_{c,max}$ and J_{max} vary with vertical depth in the canopy on account of the well-documented reduction in canopy nitrogen through the leaf profile, see [Bonan et al. 2012](#) for details). Thus, both $V_{c,max}$ and J_{max} are indexed by by Cl , ft and z according to the nitrogen decay coefficient K_n and the amount of vegetation area shading each leaf layer V_{above} ,

$$\begin{aligned} V_{c,max,Cl,ft,z} &= V_{c,max0,ft} e^{-K_{n,ft} V_{above,Cl,ft,z}}, \\ J_{max,Cl,ft,z} &= J_{max0,ft} e^{-K_{n,ft} V_{above,Cl,ft,z}}, \end{aligned}$$

where $V_{c,max,0}$ and $J_{max,0}$ are the top-of-canopy photosynthetic rates. V_{above} is the sum of exposed leaf area index ($elai_{Cl,ft,z}$) and the exposed stem area index ($esai_{Cl,ft,z}$) ($m^2 m^{-2}$). Namely,

$$V_{Cl,ft,z} = elai_{Cl,ft,z} + esai_{Cl,ft,z}.$$

The vegetation index shading a particular leaf layer in the top canopy layer is equal to

$$V_{above,Cl,ft,z} = \sum_1^z V_{Cl,ft,z} \quad \text{for } Cl = 1.$$

For lower canopy layers, the weighted average vegetation index of the canopy layer above (V_{canopy}) is added to this within-canopy shading. Thus,

$$V_{above,Cl,ft,z} = \sum_1^z V_{Cl,ft,z} + V_{canopy,Cl-1} \quad \text{for } Cl > 1,$$

where V_{canopy} is calculated as

$$V_{canopy,Cl} = \sum_{ft=1}^{npft} \sum_{z=1}^{nz(ft)} (V_{Cl,ft,z} \cdot pft_{wt,Cl,ft,1}).$$

K_n is the coefficient of nitrogen decay with canopy depth. The value of this parameter is taken from the work of [Lloyd et al. 2010](#) who determined, from 204 vertical profiles of leaf traits, that the decay rate of N through canopies of tropical rainforests was a function of the V_{cmax} at the top of the canopy. They obtain the following term to predict K_n ,

$$K_{n,ft} = e^{0.00963 V_{c,max0,ft} - 2.43},$$

where V_{cmax} is again in $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$.

Water Stress on gas exchange

¹² The top of canopy leaf photosynthetic capacity, $V_{c,max0}$, is also adjusted for the availability of water to plants as

$$V_{c,max0,25} = V_{c,max0,25} \beta_{sw},$$

where the adjusting factor β_{sw} ranges from one when the soil is wet to zero when the soil is dry. It depends on the soil water potential of each soil layer, the root distribution of the plant functional type, and a plant-dependent response to soil water stress,

$$\beta_{sw} = \sum_{j=1}^{nj} w_j r_j,$$

where w_j is a plant wilting factor for layer j and r_j is the fraction of roots in layer j . The plant wilting factor w_j is

$$w_j = \begin{cases} \frac{\psi_c - \psi_j}{\psi_c - \psi_o} \left(\frac{\theta_{sat,j} - \theta_{ice,j}}{\theta_{sat,j}} \right) & \text{for } T_i > -2C \\ 0 & \text{for } T_j \geq -2C \end{cases}$$

¹² This description relates to algorithms in the ED_btran subroutine

where ψ_i is the soil water matric potential (mm) and ψ_c and ψ_o are the soil water potential (mm) when stomata are fully closed or fully open, respectively. The term in brackets scales w_i the ratio of the effective porosity (after accounting for the ice fraction) relative to the total porosity. $w_i = 0$ when the temperature of the soil layer (T_i) is below some threshold (-2:math: $\wedge{o/C}$) or when there is no liquid water in the soil layer ($\theta_{liq,i} \leq 0$). For more details on the calculation of soil matric potential, see the CLM4.5 technical note.

Variation of water stress and water uptake within tiles

The remaining drivers of the photosynthesis model remain constant (atmospheric CO₂ and O² and canopy temperature) throughout the canopy, except for the water stress index β_{sw} . β_{sw} must be indexed by ft , because plants of differing functional types have the capacity to have varying root depth, and thus access different soil moisture profile and experience differing stress functions. Thus, the water stress function applied to gas exchange calculation is now calculated as

$$\beta_{sw,ft} = \sum_{j=1}^{nj} w_{j,ft} r_{j,ft},$$

where w_j is the water stress at each soil layer j and $r_{j,ft}$ is the root fraction of each PFT's root mass in layer j . Note that this alteration of the β_{sw} parameter also necessitates recalculation of the vertical water extraction profiles. In the original model, the fraction of extraction from each layer ($r_{e,j,patch}$) is the product of a single root distribution, because each patch only has one plant functional type. In FATES, we need to calculate a new weighted patch effective rooting depth profile $r_{e,j,patch}$ as the weighted average of the functional-type level stress functions and their relative contributions to canopy conductance. Thus for each layer j , the extraction fraction is summed over all PFTs as

$$r_{e,j,patch} = \sum_{ft=1}^{ft=npt} \frac{w_{j,ft}}{\sum_{j=1}^{nj} w_{j,ft}} \frac{G_{s,ft}}{G_{s,canopy}},$$

where nj is the number of soil layers, $G_{s,canopy}$ is the total canopy (see section 9 for details) and $G_{s,ft}$ is the canopy conductance for plant functional type ft ,

$$G_{s,ft} = \sum_1 w_{ncoh,ft} g_{s,coh} n_{coh}.$$

Aggregation of assimilated carbon into cohorts

The derivation of photosynthetic rates per leaf layer, as above, give us the estimated rate of assimilation for a unit area of leaf at a given point in the canopy in $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$. To allow the integration of these rates into fluxes per individual tree, or cohort of trees ($\text{gCO}_2 \text{ tree}^{-1} \text{ s}^{-1}$), they must be multiplied by the amount of leaf area placed in each layer by each cohort. Each cohort is described by a single functional type, ft and canopy layer C_l flag, so the problem is constrained to integrating these fluxes through the vertical profile (z).

We first make a weighted average of photosynthesis rates from sun (gpp_{sun} , $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$) and shade leaves (gpp_{shade} , $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$) as

$$\text{gpp}_{Cl,ft,z} = \text{gpp}_{sun,Cl,ft,z} f_{sun,Cl,ft,z} + \text{gpp}_{sha,Cl,ft,z} (1 - f_{sun,Cl,ft,z}).$$

The assimilation per leaf layer is then accumulated across all the leaf layers in a given cohort (coh) to give the cohort-specific gross primary productivity (GPP_{coh}),

$$GPP_{coh} = 12 \times 10^{-9} \sum_{z=1}^{nz(coh)} gpp_{Cl,ft,z} A_{crown,coh} elai_{Cl,ft,z}$$

The $elai_{l,Cl,ft,z}$ is the exposed leaf area which is present in each leaf layer in $\text{m}^2 \text{ m}^{-2}$. (For all the leaf layers that are completely occupied by a cohort, this is the same as the leaf fraction of δ_{vai}). The fluxes are converted from μmol into

mol and then multiplied by 12 (the molecular weight of carbon) to give units for GPP_{coh} of $\text{KgC cohort}^{-1} \text{ s}^{-1}$. These are integrated for each timestep to give $\text{KgC cohort}^{-1} \text{ day}^{-1}$

Table 2.44: Parameters needed for photosynthesis model.

Parameter Symbol	Parameter Name	Units	indexed by
$V_{c,max0}$	Maximum carboxylation capacity	$\mu \text{ mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$	ft
r_b	Base Rate of Respiration	$\text{gC gN}^{-1} \text{s}^{-1}$)	
q_{10}	Temp. Response of stem and root respiration		
$R_{cn,leaf,ft}$	CN ratio of leaf matter	gC/gN	ft
$R_{cn,root,ft}$	CN ratio of root matter	gC/gN	ft
f_{gr}	Growth Respiration Fraction	none	
ψ_c	Water content when stomata close	Pa	ft
ψ_o	Water content above which stomata are open	Pa	ft

2.28.8 Plant respiration

[13]_Plant respiration per individual $R_{plant,coh}$ ($\text{KgC individual}^{-1} \text{ s}^{-1}$) is the sum of two terms, growth and maintenance respiration $R_{g,coh}$ and $R_{m,coh}$

$$R_{plant} = R_{g,coh} + R_{m,coh}$$

Maintenance respiration is the sum of the respiration terms from four different plant tissues, leaf, $R_{m,leaf,coh}$, fine root $R_{m,froot,coh}$, coarse root $R_{m,croot,coh}$ and stem $R_{m,stem,coh}$, all also in ($\text{KgC individual}^{-1} \text{ s}^{-1}$).

$$R_{m,coh} = R_{m,leaf,coh} + R_{m,froot,coh} + R_{m,croot,coh} + R_{m,stem,coh}$$

To calculate canopy leaf respiration, which varies through we canopy, we first determine the top-of-canopy leaf respiration rate ($r_{m,leaf,ft,0}$, $\text{gC s}^{-1} \text{ m}^{-2}$) is calculated from a base rate of respiration per unit leaf nitrogen derived from [Ryan et al. 1991](#). The base rate for leaf respiration (r_b) is $2.525 \text{ gC/gN s}^{-1}$,

$$r_{m,leaf,ft,0} = r_b N_{a,ft} (1.5^{(25-20)/10})$$

where r_b is the base rate of metabolism ($2.525 \times 10^6 \text{ gC/gN s}^{-1}$). This base rate is adjusted assuming a Q_{10} of 1.5 to scale from the baseline of 20C to the CLM default base rate temperature of 25C. For use in the calculations of net photosynthesis and stomatal conductance, leaf respiration is converted from $\text{gC s}^{-1} \text{ m}^{-2}$, into $\mu\text{mol CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ ($/12 \cdot 10^{-6}$).

This top-of-canopy flux is scaled to account for variation in N_a through the vertical canopy, in the same manner as the $V_{c,max}$ values are scaled using V_{above} .

$$r_{leaf,Cl,ft,z} = r_{m,leaf,ft,0} e^{-K_{n,ft} V_{above,Cl,ft,z}} \beta_{ft} f(t)$$

Leaf respiration is also adjusted such that it is reduced by drought stress, β_{ft} , and canopy temperature, $f(t_{veg})$. For details of the temperature functions affecting leaf respiration see the CLM4 technical note, Section 8, Equations 8.13 and 8.14. The adjusted leaf level fluxes are scaled to individual-level ($\text{gC individual}^{-1} \text{ s}^{-1}$) in the same fashion as the GPP_{coh} calculations

$$R_{m,leaf,coh} = 12 \times 10^{-9} \sum_{z=1}^{\text{nz(coh)}} r_{leaf,Cl,ft,z} A_{\text{crown}} \text{elai}_{Cl,ft,z}$$

The stem and the coarse-root respiration terms are derived using the same base rate of respiration per unit of tissue Nitrogen.

$$R_{m,croot,coh} = 10^{-3} r_b t_c \beta_{ft} N_{\text{livecroot},coh}$$

$$R_{m,stem,coh} = 10^{-3} r_b t_c \beta_{ft} N_{\text{stem},coh}$$

Here, t_c is a temperature relationship based on a q_{10} value of 1.5, where t_v is the vegetation temperature. We use a base rate of 20 here as, again, this is the baseline temperature used by [Ryan et al. 1991](#). The 10^{-3} converts from gC individual $^{-1}$ s $^{-1}$ to KgC individual $^{-1}$ s $^{-1}$

$$t_c = q_{10}^{(t_v - 20)/10}$$

The tissue N contents for live sapwood are derived from the leaf CN ratios, and for fine roots from the root CN ratio as:

$$N_{\text{stem},coh} = \frac{B_{\text{sapwood},coh}}{R_{cn,leaf,ft}}$$

and

$$N_{\text{livecroot},coh} = \frac{B_{\text{root},coh} w_{frac,ft}}{R_{cn,root,ft}}$$

where $B_{\text{sapwood},coh}$ and $B_{\text{root},coh}$ are the biomass pools of sapwood and live root biomass respectively (KgC individual) and $w_{frac,ft}$ is the fraction of coarse root tissue in the root pool (0.5 for woody plants, 0.0 for grasses and crops). We assume here that stem CN ratio is the same as the leaf C:N ratio, for simplicity. The final maintenance respiration term is derived from the fine root respiration, which accounts for gradients of temperature in the soil profile and thus calculated for each soil layer j as follows:

$$R_{m,froot,j} = \frac{(1 - w_{frac,ft}) B_{\text{root},coh} b_r \beta_{ft}}{10^3 R_{cn,leaf,ft}} \sum_{j=1}^{nj} t_{c,soi,j} r_{i,ft,j}$$

$t_{c,soi}$ is a function of soil temperature in layer j that has the same form as that for stem respiration, but uses vertically resolved soil temperature instead of canopy temperature. In the CLM4.5, only coarse and not fine root respiration varies as a function of soil depth, and we maintain this assumption here, although it may be altered in later versions. The growth respiration, $R_{g,coh}$ is a fixed fraction f_{gr} of the carbon remaining after maintenance respiration has occurred.

$$R_{g,coh} = \max(0, GPP_{g,coh} - R_{m,coh}) f_{gr}$$

Table 2.45: Parameters needed for plant respiration model.

Parameter Symbol	Parameter Name	Units	indexed by
$-K_{n,ft}$	Rate of reduction of N through the canopy	none	•
r_b	Base Rate of Respiration	gC gN $^{-1}$ s $^{-1}$)	
q_{10}	Temp. Response of stem and root respiration		
$R_{cn,leaf,ft}$	CN ratio of leaf matter	gC/gN	ft
$R_{cn,root,ft}$	CN ratio of root matter	gC/gN	ft
f_{gr}	Growth Respiration Fraction	none	

2.28.9 Stomatal Conductance

Fundamental stomatal conductance theory

[14]_Stomatal conductance is unchanged in concept from the CLM4.5 approach. Leaf stomatal resistance is calculated from the Ball-Berry conductance model as described by [Collatz et al. 1991](#) and implemented in a global climate model by [Sellers et al. 1996](#). The model relates stomatal conductance (i.e., the inverse of resistance) to net leaf photosynthesis, scaled by the relative humidity at the leaf surface and the CO₂ concentration at the leaf surface. The primary difference between the CLM implementation and that used by [Collatz et al. 1991](#) and [Sellers et al. 1996](#) is that they used net photosynthesis (i.e., leaf photosynthesis minus leaf respiration) instead of gross photosynthesis. As implemented here, stomatal conductance equals the minimum conductance (b) when gross photosynthesis (A) is zero. Leaf stomatal resistance is

$$\frac{1}{r_s} = m_{ft} \frac{A}{c_s} \frac{e_s}{e_i} P_{atm} + b_{ft} \beta_{sw}$$

where r_s is leaf stomatal resistance (s m² μmol⁻¹), b_{ft} is a plant functional type dependent parameter equivalent to g_0 in the Ball-Berry model literature. This parameter is also scaled by the water stress index β_{sw} . Similarly, m_{ft} is the slope of the relationship between the assimilation, c_s and humidity dependant term and the stomatal conductance, and so is equivalent to the g_1 term in the stomatal literature. A is leaf photosynthesis (μmol CO₂ m⁻² s⁻¹), c_s is the CO₂ partial pressure at the leaf surface (Pa), e_s is the vapor pressure at the leaf surface (Pa), e_i is the saturation vapor pressure (Pa) inside the leaf at the vegetation temperature conductance (μmol m⁻² s⁻¹) when $A = 0$. Typical values are $m_{ft} = 9$ for C₃ plants and $m_{ft} = 4$ for C₄ plants ([Collatz et al. 1991](#), [Collatz, 1992](#), Sellers et al 1996<sellers1996>). [Sellers et al. 1996](#) used $b = 10000$ for C₃ plants and $b = 40000$ for C₄ plants. Here, b was chosen to give a maximum stomatal resistance of 20000 s m⁻¹. These terms are nevertheless plant strategy dependent, and have been found to vary widely with plant type [Medlyn et al. 2001](#).

Resistance is converted from units of s m² μ mol⁻¹ to s m⁻¹ as: 1 s m⁻¹ = 1 × 10⁻⁹ R_{gas} θ_{atm} P_{atm} (μmol⁻¹ m² s), where R_{gas} is the universal gas constant (J K⁻¹ kmol⁻¹) and θ_{atm} is the atmospheric potential temperature (K).

Resolution of stomatal conductance theory in the FATES canopy structure

The stomatal conductance is calculated, as with photosynthesis, for each canopy, PFT and leaf layer. The CLM code requires a single canopy conductance estimate to be generated from the multi-layer multi-PFT array. In previous iterations of the CLM, sun and shade-leaf specific values have been reported and then averaged by their respective leaf areas. In this version, the total canopy conductance $G_{s,canopy}$, is calculated as the sum of the cohort-level conductance values.

$$G_{s,canopy} = \sum \frac{gs_{can,coh} n_{coh}}{A_{patch}}$$

Cohort conductance is the sum of the inverse of the leaf resistances at each canopy layer ($r_{s,z}$) multiplied by the area of each cohort.

$$gs_{can,coh} = \sum_{z=1}^{z=nv,coh} \frac{A_{crown,coh}}{r_{s,cl,ft,z} + r_b}$$

Table 2.46: Parameters needed for stomatal conductance model.

Parameter Symbol	Parameter Name	Units	indexed by
b_{ft}	Slope of Ball-Berry term	none	<i>ft</i>
m_{ft}	Slope of Ball-Berry term	none	<i>ft</i>

2.28.10 Allocation and Growth

[15]_Total assimilation carbon enters the ED model each day as a cohort-specific Net Primary Productivity NPP_{coh} , which is calculated as

$$NPP_{coh} = GPP_{coh} - R_{plant,coh}$$

This flux of carbon is allocated between the demands of tissue turnover, of carbohydrate storage and of growth (increase in size of one or many plant organs). Priority is explicitly given to maintenance respiration, followed by tissue maintenance and storage, then allocation to live biomass and then to the expansion of structural and live biomass pools. All fluxes here are first converted into in KgC individual $^{-1}$ year $^{-1}$ and ultimately integrated using a timesteps of 1/365 years for each day.

Tissue maintenance demand

We calculate a ‘tissue maintenance’ flux. The magnitude of this flux is such that the quantity of biomass in each pool will remain constant, given background turnover rates. For roots, this maintenence demand is simply

$$r_{md,coh} = b_{root} \cdot \alpha_{root,ft}$$

Where $\alpha_{root,ft}$ is the root turnover rate in y^{-1} . Given that, for deciduous trees, loss of leaves is assumed to happen only one per growing season, the algorithm is dependent on phenological habit (whether or not this PFT is evergreen), thus

$$l_{md,coh} = \begin{cases} b_{leaf} \cdot \alpha_{leaf,ft} & \text{for } P_{evergreen} = 1 \\ 0 & \text{for } P_{evergreen} = 0 \end{cases}$$

Leaf litter resulting from deciduous senescence is handled in the phenology section. The total quantity of maintenance demand ($t_{md,coh}$. KgC individual y^{-1}) is therefore

$$t_{md,coh} = l_{md,coh} + r_{md,coh}$$

Allocation to storage and turnover

The model must now determine whether the NPP input is sufficient to meet the maintenance demand and keep tissue levels constant. To determine this, we introduce the idea of ‘carbon balance’ $C_{bal,coh}$ (KgC individual $^{-1}$) where

$$C_{bal,coh} = NPP_{coh} - t_{md,coh} \cdot f_{md,min,ft}$$

where $f_{md,min,ft}$ is the minimum fraction of the maintenance demand that the plant must meet each timestep, which is indexed by ft and represents a life-history-strategy decision concerning whether leaves should remain on in the case of low carbon uptake (a risky strategy) or not be replaced (a conservative strategy). Subsequently, we determine a flux to the storage pool, where the flux into the pool, as a fraction of $C_{bal,coh}$, is proportional to the discrepancy between the target pool size and the actual pool size f_{tstore} where

$$f_{tstore} = \max \left(0, \frac{b_{store}}{b_{leaf} \cdot S_{cushion}} \right)$$

The allocation to storage is a fourth power function of f_{tstore} to mimic the qualitative behaviour found for carbon allocation in arabidopsis by *Smith et al. 2007*.

$$\frac{\delta b_{store}}{\delta t} = \begin{cases} C_{bal,coh} \cdot e^{-f_{tstore}^4} & \text{for } C_{bal,coh} > 0 \\ C_{bal,coh} & \text{for } C_{bal,coh} \leq 0 \end{cases}$$

If the carbon remaining after the storage and minimum turnover fluxes have been met, the next priority is the remaining flux to leaves $t_{md} \cdot (1 - f_{md,min})$. If the quantity of carbon left ($C_{bal,coh} - \frac{\delta b_{store}}{\delta t}$) is insufficient to supply this amount of carbon, then the store of alive carbon is depleted (to represent those leaves that have fallen off and not been replaced)

$$\frac{\delta b_{alive}}{\delta t} = \begin{cases} 0 & \text{for } (C_{bal,coh} - \frac{\delta b_{store}}{\delta t}) > t_{md} \cdot (1 - f_{md,min}) \\ t_{md} \cdot (1 - f_{md,min}) - (C_{bal,coh} - \frac{\delta b_{store}}{\delta t}) & \text{for } (C_{bal,coh} - \frac{\delta b_{store}}{\delta t}) \leq t_{md} \cdot (1 - f_{md,min}) \end{cases}$$

correspondingly, the carbon left over for growth (C_{growth} : (KgC individual⁻¹ year⁻¹) is therefore

$$C_{growth} = \begin{cases} C_{bal,coh} - \frac{\delta b_{store}}{\delta t} & \text{for } (C_{bal,coh} - \frac{\delta b_{store}}{\delta t}) > 0 \\ 0 & \text{for } (C_{bal,coh} - \frac{\delta b_{store}}{\delta t}) \leq 0 \end{cases}$$

to allocate the remaining carbon (if there is any), we first ascertain whether the live biomass pool is at its target, or whether it has been depleted by previous low carbon timesteps. Thus

$$\begin{aligned} b_{alive,target} &= b_{leaf,target}(1 + f_{frla} + f_{swlhcoh}) & \text{for } S_{phen,coh} = 2 \\ b_{alive,target} &= b_{leaf,target}(f_{frla} + f_{swlhcoh}) & \text{for } S_{phen,coh} = 1 \end{aligned}$$

where the target leaf biomass $b_{leaf,target}$ ((Kg C individual⁻¹)) is the allometric relationship between dbh and leaf biomass, ameliorated by the leaf trimming fraction (see ‘control of leaf area’ below)

$$b_{leaf,target} = c_{leaf} \cdot dbh_{coh}^{e_{leaf,dbh}} \rho_{ft}^{e_{leaf,dens}} \cdot C_{trim,coh}$$

ρ_{ft} is the wood density, in g cm³.

Allocation to Seeds

The fraction remaining for growth (expansion of live and structural tissues) f_{growth} is 1 minus that allocated to seeds.

$$f_{growth,coh} = 1 - f_{seed,coh}$$

Allocation to seeds only occurs if the alive biomass is not below its target, and then is a predefined fixed fraction of the carbon remaining for growth. Allocation to clonal reproduction (primarily for grasses) occurs when \max_{dbh} is achieved.

$$f_{seed,coh} = \begin{cases} R_{frac,ft} & \text{for } \max_{dbh} < dbh_{coh} \\ (R_{frac,ft} + C_{frac,ft}) & \text{for } \max_{dbh} \geq dbh_{coh} \end{cases}$$

the total amount allocated to seed production ($p_{seed,coh}$ in KgC individual⁻¹ y⁻¹) is thus

$$p_{seed,coh} = C_{growth} \cdot f_{seed,coh}$$

Allocation to growing pools

¹⁶ The carbon is then partitioned into carbon available to grow the b_{alive} and b_{struc} pools. A fraction v_a is available to live biomass pools, and a fraction v_s is available to structural pools.

$$\frac{\delta b_{alive}}{\delta t} = C_{growth} \cdot f_{growth} v_a$$

$$\frac{\delta b_{struc}}{\delta t} = C_{growth} \cdot f_{growth} v_s$$

¹⁶ This description relates to algorithms in the ED_GrowthFunctions subroutine

If the alive biomass is lower than its ideal target, all of the available carbon is directed into that pool. Thus:

$$v_a = \begin{cases} \frac{1}{1+u} & \text{for } b_{alive} \geq b_{alive,target} \\ 1.0 & \text{for } b_{alive} < b_{alive,target} \end{cases}$$

$$v_s = \begin{cases} \frac{u}{1+u} & \text{for } b_{alive} \geq b_{alive,target} \\ 0.0 & \text{for } b_{alive} < b_{alive,target} \end{cases}$$

In this case, the division of carbon between the live and structural pools u is derived as the inverse of the sum of the rates of change in live biomass with respect to structural:

$$u = \frac{1}{\frac{\delta b_{leaf}}{\delta b_{struc}} + \frac{\delta b_{root}}{\delta b_{struc}} + \frac{\delta b_{sw}}{\delta b_{struc}}}$$

To calculate all these differentials, we first start with $\delta b_{leaf}/\delta b_{struc}$, where

$$\frac{\delta b_{leaf}}{\delta b_{struc}} = \frac{\frac{\delta dbh}{\delta b_{struc}}}{\frac{\delta dbh}{\delta b_{leaf}}}$$

The rates of change of dbh with respect to leaf and structural biomass are the differentials of the allometric equations linking these terms to each other. Hence,

$$\frac{\delta dbh}{\delta b_{leaf}} = \frac{1}{b_{trim,coh}} \cdot (e_{leaf,dbh} - 1) \exp(c_{leaf} dbh^{(e_{leaf,dbh})-1} \rho_{ft}^{e_{leaf,dens}})$$

and where $dbh_{coh} > dbh_{max}$

$$\frac{\delta b_{struc}}{\delta dbh} = e_{str,dbh} \cdot c_{str} \cdot e_{str,hite} h_{coh}^{e_{str,dbh}-1} dbh_{coh}^{e_{str,dbh}} \rho_{ft}^{e_{str,dens}}$$

If $dbh_{coh} \leq dbh_{max}$ then we must also account for allocation for growing taller as:

$$\frac{\delta b_{struc}}{\delta dbh} = \frac{\delta b_{struc}}{\delta dbh} + \frac{\delta h}{\delta dbh} \cdot \frac{\delta b_{struc}}{\delta dbh}$$

where

$$\frac{\delta h}{\delta dbh} = 1.4976 dbh_{coh}^{m_{allom}-1}$$

$$\frac{\delta dbh}{\delta b_{struc}} = \frac{1}{\frac{\delta b_{struc}}{\delta dbh}}$$

Once we have the $\delta b_{leaf}/\delta b_{struc}$, we calculate $\delta b_{root}/\delta b_{struc}$ as

$$\frac{\delta b_{root}}{\delta b_{struc}} = \frac{\delta b_{leaf}}{\delta b_{struc}} \cdot f_{frla}$$

and the sapwood differential as

$$\frac{\delta b_{sw}}{\delta b_{struc}} = f_{swh} \left(h_{coh} \frac{\delta b_{leaf}}{\delta b_{struc}} + b_{leaf,coh} \frac{\delta h}{\delta b_{struc}} \right)$$

where

$$\frac{\delta h}{\delta b_{struc}} = \frac{1}{c_{str} \times e_{str,hite} h_{coh}^{e_{str,dbh}-1} dbh_{coh}^{e_{str,dbh}} \rho_{ft}^{e_{str,dens}}}$$

In all of the above terms, height in m, dbh is in cm, and all biomass pools are in KgCm^{-2} . The allometric terms for the growth trajectory are all taken from the ED1.0 model, but could in theory be altered to accommodate alternative allometric relationships. Critically, the non-linear relationships between live and structural biomass pools are maintained in this algorithm, which diverges from the methodology currently deployed in the CLM4.5.

Integration of allocated fluxes

All of the flux calculations generate differential of the biomass state variables against time (in years). To integrate these differential rates into changes in the state variables, we use a simple simple forward Euler integration. Other methods exist (e.g. ODEINT solvers, Runge Kutta methods etc.), but they are more prone to errors that become difficult to diagnose, and the typically slow rates of change of carbon pools mean that these are less important than they might be in strongly non-linear systems (soil drainage, energy balance, etc.)

$$b_{alive,t+1} = \min \left(0, b_{alive,t} + \frac{\delta b_{alive}}{\delta t} \delta t \right)$$

$$b_{struc,t+1} = \min \left(0, b_{struc,t} + \frac{\delta b_{struc}}{\delta t} \delta t \right)$$

$$b_{store,t+1} = \min \left(0, b_{store,t} + \frac{\delta b_{store}}{\delta t} \delta t \right)$$

In this case, δt is set to be one day ($\frac{1}{365}$ years).

Table 2.47: Parameters needed for allocation model.

Parameter Symbol	Parameter Name	Units	indexed by
S	Target stored biomass as fraction of b_{leaf}	none	<i>ft</i>
f	Minimum fraction of turnover that must be met	none	<i>ft</i>
R	Fraction allocated to seeds	none	<i>ft</i>
C	Fraction allocated to clonal reproduction	none	<i>ft</i>
max_{dbh}	Diameter at which maximum height is achieved	m	<i>ft</i>
P	Does this cohort have an evergreen phenological habit?	1=yes, 0=no	<i>ft</i>

2.28.11 Control of Leaf Area Index

[17]_The leaf area A_{leaf} (m 2) of each cohort is calculated from leaf biomass $b_{leaf,coh}$ (kgC individual $^{-1}$) and specific leaf area (SLA, m 2 kg C $^{-1}$)

$$A_{leaf,coh} = b_{leaf,coh} \cdot SLA_{ft}$$

For a given tree allometry, leaf biomass is determined from basal area using the function used by *Moorcroft et al. 2001* where d_w is wood density in g cm $^{-3}$.

$$b_{leaf,coh} = c_{leaf} \cdot dbh_{coh}^{e_{leaf,dbh}} \rho_{ft}^{e_{leaf,dens}}$$

However, using this model, where leaf area and crown area are both functions of diameter, the leaf area index of each tree in a closed canopy forest is always the same (where $S_{c,patch} = S_{c,min}$, irrespective of the growth conditions). To allow greater plasticity in tree canopy structure, and for tree leaf area index to adapt to prevailing conditions, we implemented a methodology for removing those leaves in the canopy that exist in negative carbon balance. That is, their total annual assimilation rate is insufficient to pay for the turnover and maintenance costs associated with their supportive root and stem tissue, plus the costs of growing the leaf. The tissue turnover maintenance cost (KgC m $^{-2}$ y $^{-1}$) of leaf is the total maintenance demand divided by the leaf area:

$$L_{cost,coh} = \frac{t_{md,coh}}{b_{leaf,coh} \cdot SLA}$$

The net uptake for each leaf layer $U_{net,z}$ in ($\text{KgC m}^{-2} \text{ year}^{-1}$) is

$$U_{net,coh,z} = g_{coh,z} - r_{m,leaf,coh,z}$$

where g_z is the GPP of each layer of leaves in each tree ($\text{KgC m}^{-2} \text{ year}^{-1}$), $r_{m,leaf,z}$ is the rate of leaf dark respiration (also $\text{KgC m}^{-2} \text{ year}^{-1}$). We use an iterative scheme to define the cohort specific canopy trimming fraction $C_{trim,coh}$, on an annual time-step, where

$$b_{leaf,coh} = C_{trim} \times 0.0419 dbh_{coh}^{1.56} d_w^{0.55}$$

If the annual maintenance cost of the bottom layer of leaves ($\text{KgC m}^{-2} \text{ year}^{-1}$) is less than then the canopy is trimmed by an increment $\iota_l(0.01)$, which is applied until the end of next calander year. Because this is an optimality model, there is an issue of the timescale over which net assimilation is evaluated, the timescale of response, and the plasticity of plants to respond to these pressures. These properties should be investigated further in future efforts.

$$C_{trim,y+1} = \begin{cases} \max(C_{trim,y} - \iota_l, 1.0) & \text{for } (L_{cost,coh} > U_{net,coh,nz}) \\ \min(C_{trim,y} + \iota_l, L_{trim,min}) & \text{for } (L_{cost,coh} < U_{net,coh,nz}) \end{cases}$$

We impose an arbitrary minimum value on the scope of canopy trimming of $L_{trim,min}$ (0.5). If plants are able simply to drop all of their canopy in times of stress, with no consequences, then tree mortality from carbon starvation is much less likely to occur because of the greatly reduced maintenance and turnover requirements.

Table 2.48: Parameters needed for leaf area control model.

Parameter Symbol	Parameter Name	Units	indexed by
ι_l	Fraction by which leaf mass is reduced next year	none	•
$L_{trim,min}$	Minimum fraction to which leaf mass can be reduced	•	

2.28.12 Phenology

Cold Deciduous Phenology

Cold Leaf-out timing

¹⁸. The phenology model of *Botta et al. 2000* is used in FATES to determine the leaf-on timing. The Botta et al. model was verified against satellite data and is one of the only globally verified and published models of leaf-out phenology. This model differs from the phenology model in the CLM4.5. The model simulates leaf-on date as a function of the number of growing degree days (GDD), defined by the sum of mean daily temperatures (T_{day} °C) above a given threshold T_g (0 °C).

$$GDD = \sum \max(T_{day} - T_g, 0)$$

Budburst occurs when GDD exceeds a threshold (GDD_{crit}). The threshold is modulated by the number of chilling days experienced (NCD) where the mean daily temperature falls below a threshold determined by *Botta et al.*

¹⁸ This description relates to algorithms in the phenology subroutine

2000**<botta2000>** as 5°C. A greater number of chilling days means that fewer growing degree days are required before budburst:

$$GDD_{crit} = a + b e^{c \cdot NCD}$$

where $a = -68$, $b = 638$ and $c = -0.01$ *Botta et al. 2000***<botta2000>**. In the Northern Hemisphere, counting of degree days begins on 1st January, and of chilling days on 1st November. The calendar opposite of these dates is used for points in the Southern Hemisphere.

If the growing degree days exceed the critical threshold, leaf-on is triggered by a change in the gridcell phenology status flag $S_{phen,grid}$ where ‘2’ indicates that leaves should come on and ‘1’ indicates that they should fall.

$$S_{phen,grid} = 2 \quad \text{if } S_{phen,grid} = 1 \text{ and } GDD_{grid} \geq GDD_{crit}$$

Cold Leaf-off timing

The leaf-off model is taken from the Sheffield Dynamic Vegetation Model (SDGVM) and is similar to that for LPJ *Sitch et al. 2003* and IBIS *Foley et al. 1996* models. The average daily temperatures of the previous 10 day period are stored. Senescence is triggered when the number of days with an average temperature below 7.5° ($n_{colddays}$) rises above a threshold values $n_{crit,cold}$, set at 5 days.

$$S_{phen,grid} = 1 \quad \text{if } S_{phen,grid} = 2 \text{ and } n_{colddays} \geq n_{crit,cold}$$

Global implementation modifications

Because of the global implementation of the cold-deciduous phenology scheme, adjustments must be made to account for the possibility of cold-deciduous plants experiencing situations where no chilling period triggering leaf-off ever happens. If left unaccounted for, these leaves will last indefinitely, resulting in highly unrealistic behaviour. Therefore, we implement two additional rules. Firstly, if the number of days since the last senescence event was triggered is larger than 364, then leaf-off is triggered on that day. Secondly, if no chilling days have occurred during the winter accumulation period, then leaf-on is not triggered. This means that in effect, where there are no cold periods, leaves will fall off and not come back on, meaning that cold-deciduous plants can only grow in places where there is a cold season.

Further to this rule, we introduce a ‘buffer’ time periods after leaf-on of 30 days, so that cold-snap periods in the spring cannot trigger a leaf senescence. The 30 day limit is an arbitrary limit. In addition, we constrain growing degree day accumulation to the second half of the year (Jult onwards in the Northern hemisphere, or Jan-June in the Southern) and only allow GDD accumulation while the leaves are off.

Drought-deciduous Phenology: TBD

In the current version of the model, a drought deciduous algorithm exists, but is not yet operational, due to issue detected in the existing CN and soil moisture modules, which also affect the behaviour of the native ED drought deciduous model. This is a priority to address before the science tag is released.

Carbon Dynamics of deciduous plants

[19]_In the present version, leaf expansion and senescence happen over the course of a single day. This is clearly not an empirically robust representation of leaf behaviour, whereby leaf expansion occurs over a period of 10-14 days, and senescence over a similar period. This will be incorporated in later versions. When the cold or drought phenological status of the gridcell status changes ($S_{phen,grid}$) from ‘2’ to ‘1’, and the leaves are still on ($S_{phen,coh} = 2$), the leaf biomass at this timestep is ‘remembered’ by the model state variable $l_{memory,coh}$. This provides a ‘target’ biomass for

leaf onset at the beginning of the next growing season (it is a target, since depletion of stored carbon in the off season may render achieving the target impossible).

$$l_{memory,coh} = b_{leaf,coh}$$

Leaf carbon is then added to the leaf litter flux $l_{leaf,coh}$ (KgC individual $^{-1}$)

$$l_{leaf,coh} = b_{leaf,coh}$$

The alive biomass is depleted by the quantity of leaf mass lost, and the leaf biomass is set to zero

$$b_{alive,coh} = b_{alive,coh} - b_{leaf,coh}$$

$$b_{leaf,coh} = 0$$

Finally, the status $S_{phen,coh}$ is set to 1, indicating that the leaves have fallen off.

For bud burst, or leaf-on, the same occurs in reverse. If the leaves are off ($S_{phen,coh}=1$) and the phenological status triggers budburst ($S_{phen,grid}=2$) then the leaf mass is set the maximum of the leaf memory and the available store

$$b_{leaf,coh} = \max(l_{memory,coh}, b_{store,coh})$$

this amount of carbon is removed from the store

$$b_{store,coh} = b_{store,coh} - b_{leaf,coh}$$

and the new leaf biomass is added to the alive pool

$$b_{alive,coh} = b_{alive,coh} + b_{leaf,coh}$$

Lastly, the leaf memory variable is set to zero and the phenological status of the cohort back to ‘2’. No parameters are currently required for this carbon accounting scheme.

Table 2.49: Parameters needed for phenology model.

Parameter Symbol	Parameter Name	Units	indexed by
$ncrit,cold$	Threshold of cold days for senescence	none	•
T_g	Threshold for counting growing degree days	°C	

2.28.13 Seed Dynamics and Recruitment

[20]_The production of seeds and their subsequent germination is a process that must be captured explicitly or implicitly in vegetation models. FATES contains a seed bank model designed to allow the dynamics of seed production and germination to be simulated independently. In the ED1.0 model, seed recruitment occurs in the same timestep as allocation to seeds, which prohibits the survival of a viable seed bank through a period of disturbance or low productivity (winter, drought). In FATES, a plant functional type specific seed bank is tracked in each patch ($Seeds_{patch}$ KgC m $^{-2}$), whose rate of change (KgC m $^{-2}$ y $^{-1}$) is the balance of inputs, germination and decay:

$$\frac{\delta Seeds_{FT}}{\delta t} = Seed_{in,ft} - Seed_{germ,ft} - Seed_{decay,ft}$$

where $Seed_{in}$, $Seed_{germ}$ and $Seed_{decay}$ are the production, germination and decay (or onset of inviability) of seeds, all in $\text{KgC m}^{-2} \text{ year}^{-1}$.

Seeds are assumed to be distributed evenly across the site (in this version of the model), so the total input to the seed pool is therefore the sum of all of the reproductive output of all the cohorts in each patch of the correct PFT type.

$$Seed_{in,ft} = \frac{\sum_{p=1}^{n_{patch}} \sum_{i=1}^{n_{coh}} p_{seed,i} \cdot n_{coh}}{area_{site}}$$

Seed decay is the sum of all the processes that reduce the number of seeds, taken from [Lischke et al. 2006](#). Firstly, the rate at which seeds become inviable is described as a constant rate ϕ ($y:\text{math}^{\wedge}\{-1\}$) which is set to 0.51, the mean of the parameters used by [Lischke et al. 2006](#).

$$Seed_{decay,ft} = Seeds_{FT} \cdot \phi$$

The seed germination flux is also prescribed as a fraction of the existing pool (α_{sgerm}), but with a cap on maximum germination rate β_{sgerm} , to prevent excessive dominance of one plant functional type over the seed pool.

$$Seed_{germ,ft} = \max(Seeds_{FT} \cdot \alpha_{sgerm}, \beta_{sgerm})$$

Table 2.50: Parameters needed for seed model.

Parameter Symbol	Parameter Name	Units	indexed by
K_s	Maximum seed mass	kgC m^{-2}	
α_{sgerm}	Proportional germination rate	•	
β_{sgerm}	Maximum germination rate	$\text{KgC m}^{-2} \text{ y}^{-1}$	
ϕ	Decay rate of viable seeds	none	FT
$R_{frac,ft}$	Fraction of C_{bal} devoted to reproduction	none	FT

2.28.14 Litter Production and Fragmentation

The original CLM4.5 model contains streams of carbon pertaining to different chemical properties of litter (lignin, cellulose and labile streams, specifically). In FATES model, the fire simulation scheme in the SPITFIRE model requires that the model tracks the pools of litter pools that differ with respect to their propensity to burn (surface area-volume ratio, bulk density etc.). Therefore, this model contains more complexity in the representation of coarse woody debris. We also introduce the concept of 'fragmenting' pools, which are pools that can be burned, but are not available for decomposition or respiration. In this way, we can both maintain above-ground pools that affect the rate of burning, and the lag between tree mortality and availability of woody material for decomposition.

FATES recognizes four classes of litter. Above- and below-ground coarse woody debris (CWD_{AG} , CWD_{BG}) and leaf litter (l_{leaf} and fine root litter l_{root}). All pools are represented per patch, and with units of kGC m^{-2} . Further to this, CWD_{AG} , CWD_{BG} are split into four litter size classes (lsc) for the purposes of proscribing this to the SPITFIRE fire model (seed 'Fuel Load' section for more detail. 1-hour (twigs), 10-hour (small branches), 100-hour (large branches) and 1000-hour (boles or trunks). 4.5 %, 7.5%, 21 % and 67% of the woody biomass ($b_{store,coh} + b_{sw,coh}$) is partitioned into each class, respectively.

l_{leaf} and l_{root} are indexed by plant functional type (ft). The rational for indexing leaf and fine root by PFT is that leaf and fine root matter typically vary in their carbon:nitrogen ratio, whereas woody pools typically do not.

Rates of change of litter, all in $\text{kGC m}^{-2} \text{ year}^{-1}$, are calculated as

$$\begin{aligned}\frac{\delta CWD_{AG,out,lsc}}{\delta t} &= CWD_{AG,in,lsc} - CWD_{AG,out,lsc} \\ \frac{\delta CWD_{BG,out,lsc}}{\delta t} &= CWD_{BG,in,lsc} - CWD_{BG,out,lsc} \\ \frac{\delta l_{leaf,out,ft}}{\delta t} &= l_{leaf,in,ft} - l_{leaf,out,ft} \\ \frac{\delta l_{root,out,ft}}{\delta t} &= l_{root,in,ft} - l_{root,out,ft}\end{aligned}$$

Litter Inputs

[21]_Inputs into the litter pools come from tissue turnover, mortality of canopy trees, mortality of understorey trees, mortality of seeds, and leaf senescence of deciduous plants.

$$l_{leaf,in,ft} = \left(\sum_{i=1}^{n_{coh,ft}} n_{coh}(l_{md,coh} + l_{leaf,coh}) + M_{t,coh}.b_{leaf,coh} \right) / \sum_{p=1}^{n_{pat}} A_{patch}$$

where $l_{md,coh}$ is the leaf turnover rate for evergreen trees and $l_{leaf,coh}$ is the leaf loss from phenology in that timestep (KgC m^{-2}). $M_{t,coh}$ is the total mortality flux in that timestep (in individuals). For fine root input. $n_{coh,ft}$ is the number of cohorts of functional type ‘FT’ in the current patch.

$$l_{root,in,ft} = \left(\sum_{i=1}^{n_{coh,ft}} n_{coh}(r_{md,coh}) + M_{t,coh}.b_{root,coh} \right) / \sum_{p=1}^{n_{pat}} A_p$$

where $r_{md,coh}$ is the root turnover rate. For coarse woody debris input ($CWD_{AG,in,lsc}$, we first calculate the sum of the mortality $M_{t,coh}.(b_{struc,coh} + b_{sw,coh})$ and turnover $n_{coh}(w_{md,coh})$ fluxes, then separate these into size classes and above/below ground fractions using the fixed fractions assigned to each (f_{lsc} and f_{ag})

$$\begin{aligned}CWD_{AG,in,lsc} &= \left(f_{lsc}.f_{ag} \sum_{i=1}^{n_{coh,ft}} n_{coh}w_{md,coh} + M_{t,coh}.(b_{struc,coh} + b_{sw,coh}) \right) / \sum_{p=1}^{n_{pat}} A_p \\ CWD_{BG,in,lsc} &= \left(f_{lsc}.(1 - f_{ag}) \sum_{i=1}^{n_{coh,ft}} n_{coh}w_{md,coh} + M_{t,coh}.(b_{struc,coh} + b_{sw,coh}) \right) / \sum_{p=1}^{n_{pat}} A_p\end{aligned}$$

Litter Outputs

[22]_The fragmenting litter pool is available for burning but not for respiration or decomposition. Fragmentation rates are calculated according to a maximum fragmentation rate ($\alpha_{cwd,lsc}$ or α_{litter}) which is ameliorated by a temperature and water dependent scalar S_{tw} . The form of the temperature scalar is taken from the existing CLM4.5BGC decomposition cascade calculations). The water scalar is equal to the water limitation on photosynthesis (since the CLM4.5BGC water scalar pertains to the water potential of individual soil layers, which it is difficult to meaningfully average, given the non-linearities in the impact of soil moisture). The scalar code is modular, and new functions may be implemented trivially. Rate constants for the decay of the litter pools are extremely uncertain in literature, as few studies either separate litter into size classes, nor examine its decomposition under non-limiting moisture and temperature conditions. Thus, these parameters should be considered as part of sensitivity analyses of the model outputs.

$$CWD_{AG,out,lsc} = CWD_{AG,lsc}.\alpha_{cwd,lsc}.S_{tw}$$

$$\begin{aligned} CWD_{BG,out,lsc} &= CWD_{BG,lsc} \cdot \alpha_{cwd,lsc} \cdot S_{tw} \\ l_{leaf,out,ft} &= l_{leaf,ft} \cdot \alpha_{litter} \cdot S_{tw} \\ l_{root,out,ft} &= l_{root,ft} \cdot \alpha_{root,ft} \cdot S_{tw} \end{aligned}$$

Flux into decomposition cascade

[23] Upon fragmentation and release from the litter pool, carbon is transferred into the labile, lignin and cellulose decomposition pools. These pools are vertically resolved in the biogeochemistry model. The movement of carbon into each vertical layer is obviously different for above- and below-ground fragmenting pools. For each layer z and chemical litter type i , we derive a flux from ED into the decomposition cascade as $ED_{lit,i,z}$ ($\text{kGC m}^{-2} \text{s}^{-1}$)

where t_c is the time conversion factor from years to seconds, $f_{lab,l}$, $f_{cel,l}$ and $f_{lig,l}$ are the fractions of labile, cellulose and lignin in leaf litter, and $f_{lab,r}$, $f_{cel,r}$ and $f_{lig,r}$ are their counterparts for root matter. Similarly, l_{prof} , $r_{f,prof}$ and $r_{c,prof}$ are the fractions of leaf, coarse root and fine root matter that are passed into each vertical soil layer z , derived from the CLM(BGC) model.

Table 2.51: Parameters needed for litter model.

Parameter Symbol	Parameter Name	Units	indexed by
$\alpha_{ cwd,lsc }$	Maximum fragmentation rate of CWD	y^{-1}	
α_{litter}	Maximum fragmentation rate of leaf litter	y^{-1}	
α_{root}	Maximum fragmentation rate of fine root litter	y^{-1}	
$f_{lab,l}$	Fraction of leaf mass in labile carbon pool	none	
$f_{cel,l}$	Fraction of leaf mass in cellulose carbon pool	none	
$f_{lig,l}$	Fraction of leaf mass in lignin carbon pool	none	
$f_{lab,r}$	Fraction of root mass in labile carbon pool	none	
$f_{cel,r}$	Fraction of root mass in cellulose carbon pool	none	
$f_{lig,r}$	Fraction of root mass in lignin carbon pool	none	
$l_{prof,z}$	Fraction of leaf matter directed to soil layer z	none	soil layer
$r_{c,prof,z}$	Fraction of coarse root matter directed to soil layer z	none	soil layer
$r_{f,prof,z}$	Fraction of fine root matter directed to soil layer z	none	soil layer

2.28.15 Plant Mortality

Total plant mortality per cohort $M_{t,coh}$, (fraction year $^{-1}$) is simulated as the sum of four additive terms,

$$M_{t,coh} = M_{b,coh} + M_{cs,coh} + M_{hf,coh} + M_{f,coh},$$

where M_b is the background mortality that is unaccounted by any of the other mortality rates and is fixed at 0.014. M_{cs} is the carbon starvation derived mortality, which is a function of the non-structural carbon storage term $b_{store,coh}$ and the PFT-specific ‘target’ carbon storage, $l_{targ,ft}$, as follows:

$$M_{cs,coh} = \max \left(0.0, S_{m,ft} \left(0.5 - \frac{b_{store,coh}}{l_{targ,ft} b_{leaf}} \right) \right)$$

where $S_{m,ft}$ is the *stress mortality* parameter, or the fraction of trees in a landscape that die when the mean condition of a given cohort triggers mortality. This parameter is needed to scale from individual-level mortality simulation to grid-cell average conditions.

Mechanistic simulation of hydraulic failure is not undertaken on account of its mechanistic complexity (see McDowell et al. 2013<mcowell2013>‘for details). Instead, we use a proxy for hydraulic failure induced mortality ($M_{hf,coh}$) that uses a water potential threshold beyond mortality is triggered, such that the tolerance of low water potentials is a function of plant functional type (as expressed via the ψ_c parameter). For each day that the aggregate water potential falls below a threshold value, a set fraction of the trees are killed. The aggregation of soil moisture potential across the root zone is expressed using the β function. We thus determine plant mortality caused by extremely low water potentials as

$$M_{hf,coh} = \begin{cases} S_{m,ft} & \text{for } \beta_{ft} < 10^{-6} \\ 0.0 & \text{for } \beta_{ft} \geq 10^{-6}. \end{cases}$$

The threshold value of 10^{-6} represents a state where the average soil moisture potential is within 10^{-6} of the wilting point (a PFT specific parameter $\theta_{w,ft}$).

$M_{hf,coh}$ is the fire-induced mortality, as described in the fire modelling section.

Table 2.52: Parameters needed for mortality model.

Parameter Symbol	Parameter Name	Units	indexed by
$S_{m,ft}$	Stress Mortality Scaler	none	
$l_{targ,ft}$	Target carbon storage fraction	none	ft

2.28.16 Fire (SPITFIRE)

[24]_The influence of fire on vegetation is estimated using the SPITFIRE model, which has been modified for use in ED following its original implementation in the LPJ-SPITFIRE model (Thonicke et al. 2010). This model as described is substantially different from the existing CLM4.5 fire model [Li et al. 2012](#), however, further developments are intended to increase the merging of SPITFIRE’s natural vegetation fire scheme with the fire suppression, forest-clearing and peat fire estimations in the existing model. The coupling to the ED model allows fires to interact with vegetation in a size-structured manner, so small fires can burn only understorey vegetation. Also, the patch structure and representation of succession in the ED model allows the model to track the impacts of fire on different forest stands, therefore removing the problem of area-averaging implicit in area-based DGVMs. The SPITFIRE approach has also been coupled to the LPJ-GUESS individual-based model (Forrest et al. in prep) and so this is not the only implementation of this type of scheme in existence.

The SPITFIRE model operates at a daily timestep and at the patch level, meaning that different litter pools and vegetation characteristics of open and closed forests can be represented effectively (we omit the *patch* subscript throughout for simplicity).

Properties of fuel load

Many fire processes are impacted by the properties of the litter pool in the SPITFIRE model. There are one live (live grasses) and five dead fuel categories (dead leaf litter and four pools of coarse woody debris). Coarse woody debris is classified into 1h, 10h, 100h, and 1000h fuels, defined by the order of magnitude of time required for fuel to lose (or gain) 63% of the difference between its current moisture content and the equilibrium moisture content under defined atmospheric conditions. [Thonicke et al. 2010](#). For the purposes of describing the behaviour of fire, we introduce a new index ‘fuel class’ fc , the values of which correspond to each of the six possible fuel categories as follows.

<i>fc</i> index	Fuel type	Drying Time
1	dead grass	n/a
2	twigs	1h fuels
3	small branches	10h fuel
4	large branches	100h fuel
5	stems and trunks	1000h fuel
6	live grasses	n/a

Nesterov Index

Dead fuel moisture ($moist_{df,fc}$), and several other properties of fire behaviour, are a function of the ‘Nesterov Index’ (N_I) which is an accumulation over time of a function of temperature and humidity (Eqn 5, [Thonicke et al. 2010](#)).

$$N_I = \sum \max(T_d(T_d - D), 0)$$

where T_d is the daily mean temperature in °C and D is the dew point calculated as .

$$v = \frac{17.27T_d}{237.70 + T_d} + \log(RH/100)$$

$$D = \frac{237.70v}{17.27 - v}$$

where RH is the relative humidity (%).

Fuel properties

Total fuel load $F_{tot,patch}$ for a given patch is the sum of the above ground coarse woody debris and the leaf litter, plus the alive grass leaf biomass $b_{l,grass}$ multiplied by the non-mineral fraction (1- M_f).

$$F_{tot,patch} = \left(\sum_{fc=1}^{fc=5} CWD_{AG,fc} + l_{litter} + b_{l,grass} \right) (1 - M_f)$$

Many of the model behaviours are affected by the patch-level weighted average properties of the fuel load. Typically, these are calculated in the absence of 1000-h fuels because these do not contribute greatly to fire spread properties.

Dead Fuel Moisture Content

Dead fuel moisture is calculated as

$$moist_{df,fc} = e^{-\alpha_{fmc,fc} N_I}$$

where $\alpha_{fmc,fc}$ is a parameter defining how fuel moisture content varies between the first four dead fuel classes.

Live grass moisture Content

The live grass fractional moisture content($moist_{lg}$) is a function of the soil moisture content. (Equation B2 in [Thonicke et al. 2010](#))

$$moist_{lg} = \max(0.0, \frac{10}{9}\theta_{30} - \frac{1}{9})$$

where θ_{30} is the fractional moisture content of the top 30cm of soil.

Patch Fuel Moisture

The total patch fuel moisture is based on the weighted average of the different moisture contents associated with each of the different live grass and dead fuel types available (except 1000-h fuels).

$$F_{m,patch} = \sum_{fc=1}^{fc=4} \frac{F_{fc}}{F_{tot}} moist_{df,fc} + \frac{b_{l,grass}}{F_{tot}} moist_{lg}$$

Effective Fuel Moisture Content

Effective Fuel Moisture Content is used for calculations of fuel consumed, and is a function of the ratio of dead fuel moisture content $M_{df,fc}$ and the moisture of extinction factor, $m_{ef,fc}$

$$E_{moist,fc} = \frac{moist_{fc}}{m_{ef,fc}}$$

where the m_{ef} is a function of surface-area to volume ratio.

$$m_{ef,fc} = 0.524 - 0.066 \log_{10} \sigma_{fc}$$

Patch Fuel Moisture of Extinction

The patch ‘moisture of extinction’ factor (F_{mef}) is the weighted average of the m_{ef} of the different fuel classes

$$F_{mef,patch} = \sum_{fc=1}^{fc=5} \frac{F_{fc}}{F_{tot}} m_{ef,fc} + \frac{b_{l,grass}}{F_{tot}} m_{ef,grass}$$

Patch Fuel Bulk Density

The patch fuel bulk density is the weighted average of the bulk density of the different fuel classes (except 1000-h fuels).

$$F_{bd,patch} = \sum_{fc=1}^{fc=4} \frac{F_{fc}}{F_{tot}} \beta_{fuel,fc} + \frac{b_{l,grass}}{F_{tot}} \beta_{fuel,lg}$$

where $\beta_{fuel,fc}$ is the bulk density of each fuel size class (kG m^{-3})

Patch Fuel Surface Area to Volume

The patch surface area to volume ratio (F_σ) is the weighted average of the surface area to volume ratios (σ_{fuel}) of the different fuel classes (except 1000-h fuels).

$$F_\sigma = \sum_{fc=1}^{fc=4} \frac{F_{fc}}{F_{tot}} \sigma_{fuel,fc} + \frac{b_{l,grass}}{F_{tot}} \sigma_{fuel,grass}$$

Forward rate of spread

For each patch and each day, we calculate the rate of forward spread of the fire ros_f (nominally in the direction of the wind).

$$ros_f = \frac{i_r x_i (1 - \phi_w)}{F_{bd,patch} e_{ps} q_{ig}}$$

e_{ps} is the effective heating number ($e^{\frac{-4.528}{F_{\sigma,patch}}}$). q_{ig} is the heat of pre-ignition ($581 + 2594F_m$). x_i is the propagating flux calculated as (see [Thonicke et al. 2010](#) Appendix A).

$$x_i = \begin{cases} 0.0 & \text{for } F_{\sigma,patch} < 0.00001 \\ \frac{e^{0.792+3.7597F_{\sigma,patch}^{0.5}}(\frac{F_{bd,patch}}{p_d}+0.1)}{192+7.9095F_{\sigma,patch}} & \text{for } F_{\sigma,patch} \geq 0.00001 \end{cases}$$

ϕ_w is the influence of windspeed on rate of spread.

$$\phi_w = cb_w^b \cdot \beta^{-e}$$

Where b , c and e are all functions of surface-area-volume ratio $F_{\sigma,patch}$: $b = 0.15988F_{\sigma,patch}^{0.54}$, $c = 7.47e^{-0.8711F_{\sigma,patch}^{0.55}}$, $e = 0.715e^{-0.01094F_{\sigma,patch}}$. $b_w = 196.86W$ where W is the the windspeed in ms^{-1} , and $\beta = \frac{F_{ba}/p_d}{0.200395F_{\sigma,patch}^{-0.8189}}$ where p_d is the particle density (513).

i_r is the reaction intensity, calculated using the following set of expressions (from [Thonicke et al. 2010](#) Appendix A):

$$\begin{aligned} i_r &= \Gamma_{opt} F_{tot} H d_{moist} d_{miner} \\ d_{moist} &= \max(0.0, (1 - 2.59m_w + 5.11m_w^2 - 3.52m_w^3)) \\ m_w &= \frac{F_{m,patch}}{F_{mef,patch}} \\ \Gamma_{opt} &= \Gamma_{max} \beta^a \lambda \\ \Gamma_{max} &= \frac{1}{0.0591 + 2.926F_{\sigma,patch}^{-1.5}} \\ \lambda &= e^{a(1-\beta)} \\ a &= 8.9033F_{\sigma,patch}^{-0.7913} \end{aligned}$$

Γ_{opt} is the residence time of the fire, and d_{miner} is the mineral damping coefficient (=0.174:math:S_e^{-0.19}), where S_e is 0.01 and so $= d_{miner}$ 0.41739).

Fuel Consumption

The fuel consumption (fraction of biomass pools) of each dead biomass pool in the area affected by fire on a given day ($f_{c,dead,fc}$) is a function of effective fuel moisture $E_{moist,fc}$ and size class fc (Eqn B1, B4 and B5, [Thonicke et al. 2010](#)). The fraction of each fuel class that is consumed decreases as its moisture content relative to its moisture of extinction ($E_{moist,fc}$) increases.

$$f_{cdead,fc} = \max \left(0, \min(1, m_{int,mc,fc} - m_{slope,mc,fc} E_{moist,fc}) \right)$$

m_{int} and m_{slope} are parameters, the value of which is modulated by both size class fc and by the effective fuel moisture class mc , defined by $E_{moist,fc}$. m_{int} and m_{slope} are defined for low-, mid-, and high-moisture conditions, the boundaries of which are also functions of the litter size class following [Peterson and Ryan 1986](#) (page 802). The fuel burned, $f_{cground,fc}$ ($\text{Kg m}^{-2} \text{ day}^{-1}$) is calculated from $f_{cdead,fc}$ for each fuel class:

$$f_{cground,fc} = f_{c,dead,fc}(1 - M_f) \frac{F_{fc}}{0.45}$$

Where 0.45 converts from carbon to biomass. The total fuel consumption, $f_{ctot,patch}$ (Kg m⁻²), used to calculate fire intensity, is then given by

$$f_{ctot,patch} = \sum_{fc=1}^{fc=4} f_{c,ground,fc} + f_{c,ground,lgrass}$$

There is no contribution from the 1000 hour fuels to the patch-level $f_{ctot,patch}$ used in the fire intensity calculation.

Fire Intensity

Fire intensity at the front of the burning area ($I_{surface}$, kW m⁻²) is a function of the total fuel consumed ($f_{ctot,patch}$) and the rate of spread at the front of the fire, ros_f (m min⁻¹) (Eqn 15 [Thonicke et al. 2010](#))

$$I_{surface} = \frac{0.001}{60} f_{energy} f_{ctot,patch} ros_f$$

where f_{energy} is the energy content of fuel (kJ/kG - the same for all fuel classes). Fire intensity is used to define whether an ignition is successful. If the fire intensity is greater than 50kw/m then the ignition is successful.

Fire Duration

Fire duration is a function of the fire danger index with a maximum length of $F_{dur,max}$ (240 minutes in [Thonicke et al. 2010](#) Eqn 14, derived from Canadian Forest Fire Behaviour Predictions Systems)

$$D_f = \min\left(F_{dur,max}, \frac{F_{dur,max}}{1 + F_{dur,max} e^{-11.06 fdi}}\right)$$

Fire Danger Index

Fire danger index (fdi) is a representation of the effect of meteorological conditions on the likelihood of a fire. It is calculated for each gridcell as a function of the Nesterov Index . fdi is calculated from NI as

$$fdi = 1 - e^{\alpha NI}$$

where $\alpha = 0.00037$ following [Venevsky et al. 2002](#).

Area Burned

Total area burnt is assumed to be in the shape of an ellipse, whose major axis f_{length} (m) is determined by the forward and backward rates of spread (ros_f and ros_b respectively).

$$f_{length} = F_d(ros_b + ros_f)$$

ros_b is a function of ros_f and windspeed (Eqn 10 [Thonicke et al. 2010](#))

$$ros_b = ros_f e^{-0.72W}$$

The minor axis to major axis ratio l_b of the ellipse is determined by the windspeed. If the windspeed (W) is less than 16.67 ms⁻¹ then $l_b = 1$. Otherwise (Eqn 12 and 13, [Thonicke et al. 2010](#))

$$l_b = \min\left(8, f_{tree}(1.0 + 8.729(1.0 - e^{-0.108W})^{2.155}) + (f_{grass}(1.1 + 3.6W^{0.0464}))\right)$$

f_{grass} and f_{tree} are the fractions of the patch surface covered by grass and trees respectively.

The total area burned (A_{burn} in m^2) is therefore (Eqn 11, *Thonicke et al. 2010*)

$$A_{burn} = \frac{n_f \frac{3.1416}{4l_b} (f_{length}^2))}{10000}$$

where n_f is the number of fires.

Crown Damage

c_k is the fraction of the crown which is consumed by the fire. This is calculated from scorch height H_s , tree height h and the crown fraction parameter F_{crown} (Eqn 17 *Thonicke et al. 2010*):

$$c_k = \begin{cases} 0 & \text{for } H_s < (h - hF_{crown}) \\ 1 - \frac{h - H_s}{h - F_{crown}} & \text{for } h > H_s > (h - hF_{crown}) \\ 1 & \text{for } H_s > h \end{cases}$$

The scorch height H_s (m) is a function of the fire intensity, following *Byram, 1959*, and is proportional to a plant functional type specific parameter $\alpha_{s,ft}$ (Eqn 16 *Thonicke et al. 2010*):

$$H_s = \sum_{FT=1}^{NPFT} \alpha_{s,p} \cdot f_{biomass,ft} I_{surface}^{0.667}$$

where $f_{biomass,ft}$ is the fraction of the above-ground biomass in each plant functional type.

Cambial Damage and Kill

The cambial kill is a function of the fuel consumed $f_{c,tot}$, the bark thickness t_b , and τ_l , the duration of cambial heating (minutes) (Eqn 8, *Peterson and Ryan 1986*):

$$\tau_l = \sum_{fc=1}^{fc=5} 39.4 F_{p,c} \frac{10000}{0.45} (1 - (1 - f_{c,dead,fc})^{0.5})$$

Bark thickness is a linear function of tree diameter dbh_{coh} , defined by PFT-specific parameters $\beta_{1,bt}$ and $\beta_{2,bt}$ (Eqn 21 *Thonicke et al. 2010*):

$$t_{b,coh} = \beta_{1,bt,ft} + \beta_{2,bt,ft} dbh_{coh}$$

The critical time for cambial kill, τ_c (minutes) is given as (Eqn 20 *Thonicke et al. 2010*):

$$\tau_c = 2.9 t_b^2$$

The mortality rate caused by cambial heating τ_{pm} of trees within the area affected by fire is a function of the ratio between τ_l and τ_c (Eqn 19, *Thonicke et al. 2010*):

$$\tau_{pm} = \begin{cases} 1.0 & \text{for } \tau_l/\tau_c \geq 2.0 \\ 0.563(\tau_l/\tau_c)) - 0.125 & \text{for } 2.0 > \tau_l/\tau_c \geq 0.22 \\ 0.0 & \text{for } \tau_l/\tau_c < 0.22 \end{cases}$$

Table 2.53: Parameters needed for fire model.

Parameter Symbol	Parameter Name	Units	indexed by
$\beta_{1,bt}$	Intercept of bark thickness function	mm	<i>FT</i>
$\beta_{2,bt}$	Slope of bark thickness function	mm cm ⁻¹	<i>FT</i>
F_{crown}	Ratio of crown height to total height	none	<i>FT</i>
α_{fmc}	Fuel moisture parameter	°C ⁻²	<i>fc</i>
β_{fuel}	Fuel Bulk Density	kG m ⁻³	<i>fc</i>
$\sigma_{fuel,fc}$	Surface area to volume ratio	cm ⁻¹	<i>fc</i>
m_{int}	Intercept of fuel burned	none	<i>fc</i> , moisture class
m_{slope}	Slope of fuel burned	none	<i>fc</i> , moisture class
M_f	Fuel Mineral Fraction		
$F_{dur,max}$	Maximum Duration of Fire	Minutes	
f_{energy}	Energy content of fuel	kJ/kG	
α_s	Flame height parameter		<i>FT</i>

2.29 Biogenic Volatile Organic Compounds (BVOCs)

This chapter briefly describes the biogenic volatile organic compound (BVOC) emissions model implemented in CLM. The CLM3 version (Levis et al. 2003; Oleson et al. 2004) was based on Guenther et al. (1995). Heald et al. (2008) updated this scheme in CLM4 based on Guenther et al. (2006). The current version was implemented in CLM4.5 and is based on MEGAN2.1 discussed in detail in Guenther et al. (2012). This update of MEGAN incorporates four main features: 1) expansion to 147 chemical compounds, 2) the treatment of the light-dependent fraction (LDF) for each compound, 3) inclusion of the inhibition of isoprene emission by atmospheric CO₂ and 4) emission factors mapped to the specific PFTs of the CLM.

MEGAN2.1 now describes the emissions of speciated monoterpenes, sesquiterpenes, oxygenated VOCs as well as isoprene. A flexible scheme has been implemented in the CLM to specify a subset of emissions. This allows for additional flexibility in grouping chemical compounds to form the lumped species frequently used in atmospheric chemistry. The mapping or grouping is therefore defined through a namelist parameter in `drv flds_in`, e.g. `megan_specifier = 'ISOP = isoprene', 'BIGALK = pentane + hexane + heptane + tricyclene'`.

Terrestrial BVOC emissions from plants to the atmosphere are expressed as a flux, F_i ($\mu\text{ g C m}^{-2}$ ground area h^{-1}), for emission of chemical compound i

$$F_i = \gamma_i \rho \sum_j \varepsilon_{i,j} (\text{wt})_j \quad (2.981)$$

where γ_i is the emission activity factor accounting for responses to meteorological and phenological conditions, ρ is the canopy loss and production factor also known as escape efficiency (set to 1), and $\varepsilon_{i,j}$ ($\mu\text{ g C m}^{-2}$ ground area h^{-1}) is the emission factor at standard conditions of light, temperature, and leaf area for plant functional type j with fractional coverage $(\text{wt})_j$ (Guenther et al. 2012). The emission activity factor γ_i depends on plant functional type, temperature, LAI, leaf age, and soil moisture (Guenther et al. 2012). For isoprene only, the effect of CO₂ inhibition is now included as described by Heald et al. (2009). Previously, only isoprene was treated as a light-dependent emission. In MEGAN2.1, each chemical compound is assigned a LDF (ranging from 1.0 for isoprene to 0.2 for some monoterpenes, VOCs and acetone). The activity factor for the light response of emissions is therefore estimated as:

$$\gamma_{P,i} = (1 - LDF_i) + \gamma_{P,LDF} LDF_i \quad (2.982)$$

where the LDF activity factor ($\gamma_{P,LDF}$) is specified as a function of PAR as in previous versions of MEGAN.

The values for each emission factor $\epsilon_{i,j}$ are now available for each of the plant functional types in the CLM and each chemical compound. This information is distributed through an external file, allowing for more frequent and easier updates.

2.30 Dust Model

Atmospheric dust is mobilized from the land by wind in the CLM. The most important factors determining soil erodibility and dust emission include the wind friction speed, the vegetation cover, and the soil moisture. The CLM dust mobilization scheme ([Mahowald et al. 2006](#)) accounts for these factors based on the DEAD (Dust Entrainment and Deposition) model of [Zender et al. \(2003\)](#). Please refer to the [Zender et al. \(2003\)](#) article for additional information regarding the equations presented in this section.

The total vertical mass flux of dust, F_j ($\text{kg m}^{-2} \text{s}^{-1}$), from the ground into transport bin j is given by

$$F_j = TS f_m \alpha Q_s \sum_{i=1}^I M_{i,j} \quad (2.983)$$

where T is a global factor that compensates for the DEAD model's sensitivity to horizontal and temporal resolution and equals 5×10^{-4} in the CLM instead of 7×10^{-4} in [Zender et al. \(2003\)](#). S is the source erodibility factor set to 1 in the CLM and serves as a place holder at this time.

The grid cell fraction of exposed bare soil suitable for dust mobilization f_m is given by

$$f_m = (1 - f_{lake} - f_{wetl}) (1 - f_{sno}) (1 - f_v) \frac{w_{liq,1}}{w_{liq,1} + w_{ice,1}} \quad (2.984)$$

where f_{lake} and f_{wetl} and f_{sno} are the CLM grid cell fractions of lake and wetland (section 2.2.3) and snow cover (section 2.8.1), all ranging from zero to one. Not mentioned by [Zender et al. \(2003\)](#), $w_{liq,1}$ and $w_{ice,1}$ are the CLM top soil layer liquid water and ice contents (mm) entered as a ratio expressing the decreasing ability of dust to mobilize from increasingly frozen soil. The grid cell fraction of vegetation cover, f_v , is defined as

$$0 \leq f_v = \frac{L + S}{(L + S)_t} \leq 1 \quad \text{where } (L + S)_t = 0.3 \text{ m}^2 \text{m}^{-2} \quad (2.985)$$

where equation applies only for dust mobilization and is not related to the plant functional type fractions prescribed from the CLM input data or simulated by the CLM dynamic vegetation model (Chapter 22). L and S are the CLM leaf and stem area index values ($\text{m}^2 \text{m}^{-2}$) averaged at the land unit level so as to include all the pfts and the bare ground present in a vegetated land unit. L and S may be prescribed from the CLM input data (section 2.2.1) or simulated by the CLM biogeochemistry model (Chapter 2.20).

The sandblasting mass efficiency α (m^{-1}) is calculated as

$$\alpha = 100 e^{(13.4 M_{clay} - 6.0) \ln 10} \begin{cases} M_{clay} = \%clay \times 0.01 & 0 \leq \%clay \leq 20 \\ M_{clay} = 20 \times 0.01 & 20 < \%clay \leq 100 \end{cases} \quad (2.986)$$

where M_{clay} is the mass fraction of clay particles in the soil and $\%clay$ is determined from the surface dataset (section 2.2.3). $M_{clay} = 0$ corresponds to sand and $M_{clay} = 0.2$ to sandy loam.

Q_s is the total horizontally saltating mass flux ($\text{kg m}^{-1} \text{s}^{-1}$) of “large” particles (Table 2.54), also referred to as the vertically integrated streamwise mass flux

$$Q_s = \begin{cases} \frac{c_s \rho_{atm} u_{*s}^3}{g} \left(1 - \frac{u_{*t}}{u_{*s}}\right) \left(1 + \frac{u_{*t}}{u_{*s}}\right)^2 & \text{for } u_{*t} < u_{*s} \\ 0 & \text{for } u_{*t} \geq u_{*s} \end{cases} \quad (2.987)$$

where the saltation constant c_s equals 2.61 and ρ_{atm} is the atmospheric density (kg m^{-3}) (Table 2.4), g the acceleration of gravity (m s^{-2}) (Table 2.7). The threshold wind friction speed for saltation u_{*t} (m s^{-1}) is

$$u_{*t} = f_z \left[Re_{*t}^f \rho_{osp} g D_{osp} \left(1 + \frac{6 \times 10^{-7}}{\rho_{osp} g D_{osp}^{2.5}}\right) \right]^{\frac{1}{2}} \rho_{atm}^{-\frac{1}{2}} f_w \quad (2.988)$$

where f_z is a factor dependent on surface roughness but set to 1 as a place holder for now, ρ_{osp} and D_{osp} are the density (2650 kg m^{-3}) and diameter ($75 \times 10^{-6} \text{ m}$) of optimal saltation particles, and f_w is a factor dependent on soil moisture:

$$f_w = \begin{cases} 1 & \text{for } w \leq w_t \\ \sqrt{1 + 1.21 [100(w - w_t)]^{0.68}} & \text{for } w > w_t \end{cases} \quad (2.989)$$

where

$$w_t = a (0.17 M_{clay} + 0.14 M_{clay}^2) \quad 0 \leq M_{clay} = \%clay \times 0.01 \leq 1 \quad (2.990)$$

and

$$w = \frac{\theta_1 \rho_{liq}}{\rho_{d,1}} \quad (2.991)$$

where $a = M_{clay}^{-1}$ for tuning purposes, θ_1 is the volumetric soil moisture in the top soil layer ($\text{m}^3 \text{m}^{-3}$) (section 2.7.3), ρ_{liq} is the density of liquid water (kg m^{-3}) (Table 2.7), and $\rho_{d,1}$ is the bulk density of soil in the top soil layer (kg m^{-3}) defined as in section 2.6.3 rather than as in Zender et al. (2003). Re_{*t}^f from equation is the threshold friction Reynolds factor

$$Re_{*t}^f = \begin{cases} \frac{0.1291^2}{-1+1.928Re_{*t}} & \text{for } 0.03 \leq Re_{*t} \leq 10 \\ 0.12^2 (1 - 0.0858 e^{-0.0617(Re_{*t}-10)})^2 & \text{for } Re_{*t} > 10 \end{cases} \quad (2.992)$$

and Re_{*t} is the threshold friction Reynolds number approximation for optimally sized particles

$$Re_{*t} = 0.38 + 1331 (100 D_{osp})^{1.56} \quad (2.993)$$

In (2.987), u_{*s} is defined as the wind friction speed (m s^{-1}) accounting for the Owen effect (Owen 1964)

$$u_s = \begin{cases} u & \text{for } U_{10} < U_{10,t} \\ u_* + 0.003 (U_{10} - U_{10,t})^2 & \text{for } U_{10} \geq U_{10,t} \end{cases} \quad (2.994)$$

where u_* is the CLM wind friction speed (m s^{-1}), also known as friction velocity (section 2.5.1), U_{10} is the 10-m wind speed (m s^{-1}) calculated as the wind speed at the top of the canopy in section 4.3 of Bonan (1996) but here for 10 m above the ground, and $U_{10,t}$ is the threshold wind speed at 10 m (m s^{-1})

$$U_{10,t} = u_{*t} \frac{U_{10}}{u_*} \quad (2.995)$$

In equation we sum $M_{i,j}$ over $I = 3$ source modes i where $M_{i,j}$ is the mass fraction of each source mode i carried in each of :math:`'J=4'` transport bins j

$$M_{i,j} = \frac{m_i}{2} \left[\operatorname{erf} \left(\frac{\ln \frac{D_{j,\max}}{\tilde{D}_{v,i}}}{\sqrt{2} \ln \sigma_{g,i}} \right) - \operatorname{erf} \left(\frac{\ln \frac{D_{j,\min}}{\tilde{D}_{v,i}}}{\sqrt{2} \ln \sigma_{g,i}} \right) \right] \quad (2.996)$$

where m_i , $\tilde{D}_{v,i}$, and $\sigma_{g,i}$ are the mass fraction, mass median diameter, and geometric standard deviation assigned to each particle source mode i (Table 2.54), while $D_{j,\min}$ and $D_{j,\max}$ are the minimum and maximum diameters (m) in each transport bin j (Table 2.55).

Table 2.54: Mass fraction m_i , mass median diameter $\tilde{D}_{v,i}$, and geometric standard deviation $\sigma_{g,i}$, per dust source mode i

i	m_i (fraction)	$\tilde{D}_{v,i}$ (m)	$\sigma_{g,i}$
1	0.036	0.832×10^{-6}	2.1
2	0.957	4.820×10^{-6}	1.9
3	0.007	19.38×10^{-6}	1.6

Table 2.55: Minimum and maximum particle diameters in each dust transport bin j

j	$D_{j,\min} (\text{m})$	$D_{j,\max} (\text{m})$
1	0.1×10^{-6}	1.0×10^{-6}
2	1.0×10^{-6}	2.5×10^{-6}
3	2.5×10^{-6}	5.0×10^{-6}
4	5.0×10^{-6}	10.0×10^{-6}

2.31 Carbon Isotopes

CLM includes a fully prognostic representation of the fluxes, storage, and isotopic discrimination of the carbon isotopes ^{13}C and ^{14}C . The implementation of the C isotopes capability takes advantage of the CLM hierarchical data structures, replicating the carbon state and flux variable structures at the column and PFT level to track total carbon and both C isotopes separately (see description of data structure hierarchy in Chapter 2). For the most part, fluxes and associated updates to carbon state variables for ^{13}C are calculated directly from the corresponding total C fluxes. Separate calculations are required in a few special cases, such as where isotopic discrimination occurs, or where the necessary isotopic ratios are undefined. The general approach for ^{13}C flux and state variable calculation is described here, followed by a description of all the places where special calculations are required.

2.31.1 General Form for Calculating ^{13}C and ^{14}C Flux

In general, the flux of ^{13}C and corresponding to a given flux of total C (CF_{13C} and CF_{totC} , respectively) is determined by CF_{totC} , the masses of ^{13}C and total C in the upstream pools (CS_{13C_up} and CS_{totC_up} , respectively, i.e. the pools from which the fluxes of ^{13}C and total C originate), and a fractionation factor, f_{frac} :

$$CF_{13C} = \begin{cases} CF_{totC} \frac{CS_{13C_up}}{CS_{totC_up}} f_{frac} & \text{for } CS_{totC} \neq 0 \\ 0 & \text{for } CS_{totC} = 0 \end{cases} \quad (2.997)$$

If the $f_{frac} = 1.0$ (no fractionation), then the fluxes CF_{13C} and CF_{totC} will be in simple proportion to the masses CS_{13C_up} and CS_{totC_up} . Values of $f_{frac} < 1.0$ indicate a discrimination against the heavier isotope (^{13}C) in the flux-generating process, while $f_{frac} > 1.0$ would indicate a preference for the heavier isotope. Currently, in all cases where Eq. is used to calculate a ^{13}C flux, f_{frac} is set to 1.0.

For ^{14}C , no fractionation is used in either the initial photosynthetic step, nor in subsequent fluxes from upstream to downstream pools; as discussed below, this is because observations of ^{14}C are typically described in units that implicitly correct out the fractionation of ^{14}C by referencing them to ^{13}C ratios.

2.31.2 Isotope Symbols, Units, and Reference Standards

Carbon has two primary stable isotopes, ^{12}C and ^{13}C . ^{12}C is the most abundant, comprising about 99% of all carbon. The isotope ratio of a compound, R_A , is the mass ratio of the rare isotope to the abundant isotope

$$R_A = \frac{^{13}\text{C}_A}{^{12}\text{C}_A}. \quad (2.998)$$

Carbon isotope ratios are often expressed using delta notation, δ . The $\delta^{13}\text{C}$ value of a compound A, $\delta^{13}\text{C}_A$, is the difference between the isotope ratio of the compound, R_A , and that of the Pee Dee Belemnite standard, R_{PDB} , in parts per thousand

$$\delta^{13}\text{C}_A = \left(\frac{R_A}{R_{PDB}} - 1 \right) \times 1000 \quad (2.999)$$

where $R_{PDB} = 0.0112372$, and units of δ are per mil (\textperthousand).

Isotopic fractionation can be expressed in several ways. One expression of the fractionation factor is with alpha (α) notation. For example, the equilibrium fractionation between two reservoirs A and B can be written as:

$$\alpha_{A-B} = \frac{R_A}{R_B} = \frac{\delta_A + 1000}{\delta_B + 1000}. \quad (2.1000)$$

This can also be expressed using epsilon notation (ϵ), where

$$\alpha_{A-B} = \frac{\epsilon_{A-B}}{1000} + 1 \quad (2.1001)$$

In other words, if $\epsilon_{A-B} = 4.4 \text{\textperthousand}$, then $\alpha_{A-B} = 1.0044$.

In addition to the stable isotopes ^{12}C and ^{13}C , the unstable isotope ^{14}C is included in CLM. ^{14}C can also be described using the delta notation:

$$\delta^{14}\text{C} = \left(\frac{A_s}{A_{abs}} - 1 \right) \times 1000 \quad (2.1002)$$

However, observations of ^{14}C are typically fractionation-corrected using the following notation:

$$\Delta^{14}\text{C} = 1000 \times \left(\left(1 + \frac{\delta^{14}\text{C}}{1000} \right) \frac{0.975^2}{\left(1 + \frac{\delta^{13}\text{C}}{1000} \right)^2} - 1 \right) \quad (2.1003)$$

where $\delta^{14}\text{C}$ is the measured isotopic fraction and $\Delta^{14}\text{C}$ corrects for mass-dependent isotopic fractionation processes (assumed to be 0.975 for fractionation of ^{13}C by photosynthesis). CLM assumes a background preindustrial atmospheric $^{14}\text{C}/\text{C}$ ratio of 10^{-12} , which is used for A_{abs} . For the reference standard A_{abs} , which is a plant tissue and has a $\delta^{13}\text{C}$ value of $-25 \text{\textperthousand}$ due to photosynthetic discrimination, $\delta^{14}\text{C} = \Delta^{14}\text{C}$. For CLM, in order to use the ^{14}C model independently of the ^{13}C model, for the ^{14}C calculations, this fractionation is set to zero, such that the 0.975 term becomes 1, the $\delta^{13}\text{C}$ term (for the calculation of $\delta^{14}\text{C}$ only) becomes 0, and thus $\delta^{14}\text{C} = \Delta^{14}\text{C}$.

2.31.3 Carbon Isotope Discrimination During Photosynthesis

Photosynthesis is modeled in CLM as a two-step process: diffusion of CO_2 into the stomatal cavity, followed by enzymatic fixation (Chapter 2.9). Each step is associated with a kinetic isotope effect. The kinetic isotope effect during diffusion of CO_2 through the stomatal opening is $4.4 \text{\textperthousand}$. The kinetic isotope effect during fixation of CO_2 with Rubisco is $\sim 30 \text{\textperthousand}$; however, since about 5-10% of carbon in C3 plants reacts with phosphoenolpyruvate carboxylase (PEPC) (Melzer and O'Leary, 1987), the net kinetic isotope effect during fixation is $\sim 27 \text{\textperthousand}$ for C3 plants. In C4 plant photosynthesis, only the diffusion effect is important. The fractionation factor equations for C3 and C4 plants are given below:

For C4 plants,

$$\alpha_{psn} = 1 + \frac{4.4}{1000} \quad (2.1004)$$

For C3 plants,

$$\alpha_{psn} = 1 + \frac{4.4 + 22.6 \frac{c_i^*}{p\text{CO}_2}}{1000} \quad (2.1005)$$

where α_{psn} is the fractionation factor, and c_i^* and $p\text{CO}_2$ are the revised intracellular and atmospheric CO_2 partial pressure, respectively.

As can be seen from the above equation, kinetic isotope effect during fixation of CO_2 is dependent on the intracellular CO_2 concentration, which in turn depends on the net carbon assimilation. That is calculated during the photosynthesis calculation as follows:

$$c_i = p\text{CO}_2 - a_n p \frac{(1.4g_s) + (1.6g_b)}{g_b g_s} \quad (2.1006)$$

where a_n is net carbon assimilation during photosynthesis, p is atmospheric pressure, g_b is leaf boundary layer conductance, and g_s is leaf stomatal conductance.

Isotopic fractionation code is compatible with multi-layered canopy parameterization; i.e., it is possible to calculate varying discrimination rates for each layer of a multi-layered canopy. However, as with the rest of the photosynthesis model, the number of canopy layers is currently set to one by default.

2.31.4 ^{14}C radioactive decay and historical atmospheric ^{14}C and ^{13}C concentrations

In the preindustrial biosphere, radioactive decay of ^{14}C in carbon pools allows dating of long-term age since photosynthetic uptake; while over the 20th century, radiocarbon in the atmosphere was first diluted by radiocarbon-free fossil fuels and then enriched by aboveground thermonuclear testing to approximately double its long-term mean concentration. CLM includes both of these processes to allow comparison of carbon that may vary on multiple timescales with observed values.

For radioactive decay, at each timestep all ^{14}C pools are reduced at a rate of $-\log/\tau$, where τ is the half-life (Libby half-life value of 5568 years). In order to rapidly equilibrate the long-lived pools during accelerated decomposition spinup, the radioactive decay of the accelerated pools is also accelerated by the same degree as the decomposition, such that the ^{14}C value of these pools is in equilibrium when taken out of the spinup mode.

For variation of atmospheric ^{14}C and ^{13}C over the historical period, $\Delta^{14}\text{C}$ and $\Delta^{13}\text{C}$ values can be set to either fixed concentrations or time-varying concentrations read in from a file. A default file is provided that spans the historical period (*Graven et al., 2017*). For $\Delta^{14}\text{C}$, values are provided and read in for three latitude bands (30 °N-90 °N, 30 °S-30 °N, and 30 °S-90 °S).

2.32 Land-Only Mode

In land-only mode (uncoupled to an atmospheric model), the atmospheric forcing required by CLM (Table 2.4) is supplied by observed datasets. The standard forcing provided with the model is a 110-year (1901-2010) dataset provided by the Global Soil Wetness Project (GSWP3; NEED A REFERENCE). The GSWP3 dataset has a spatial resolution of 0.5° X 0.5° and a temporal resolution of three hours.

An alternative forcing dataset is also available, CRUNCEP, a 110-year (1901-2010) dataset (CRUNCEP; *Viovy 2011*) that is a combination of two existing datasets; the CRU TS3.2 0.5° X 0.5° monthly data covering the period 1901 to 2002 (*Mitchell and Jones 2005*) and the NCEP reanalysis 2.5° X 2.5° 6-hourly data covering the period 1948 to 2010. The CRUNCEP dataset has been used to force CLM for studies of vegetation growth, evapotranspiration, and gross primary production (*Mao et al. 2012, Mao et al. 2013, Shi et al. 2013*) and for the TRENDY (trends in net land-atmosphere carbon exchange over the period 1980-2010) project (*Piao et al. 2012*). Version 7 is available here (*Viovy 2011*).

Here, the GSWP3 dataset, which does not include data for particular fields over oceans, lakes, and Antarctica is modified. This missing data is filled with *Qian et al. (2006)* data from 1948 that is interpolated by the data atmosphere model to the 0.5° GSWP3 grid. This allows the model to be run over Antarctica and ensures data is available along coastlines regardless of model resolution.

The forcing data is ingested into a data atmosphere model in three “streams”; precipitation (P) (mm s⁻¹), solar radiation (S_{atm}) (W m⁻²), and four other fields [atmospheric pressure P_{atm} (Pa), atmospheric specific humidity q_{atm} (kg kg⁻¹), atmospheric temperature T_{atm} (K), and atmospheric wind W_{atm} (m s⁻¹)]. These are separate streams because they are handled differently according to the type of field. In the GSWP3 dataset, the precipitation stream is provided at three hour intervals and the data atmosphere model prescribes the same precipitation rate for each model time step within the three hour period. The four fields that are grouped together in another stream (pressure, humidity, temperature, and wind) are provided at three hour intervals and the data atmosphere model linearly interpolates these fields to the time step of the model.

The total solar radiation is also provided at three hour intervals. The data is fit to the model time step using a diurnal function that depends on the cosine of the solar zenith angle μ to provide a smoother diurnal cycle of solar radiation and to ensure that all of the solar radiation supplied by the three-hourly forcing data is actually used. The solar radiation at model time step t_M is

$$\begin{aligned} S_{atm}(t_M) &= \frac{\frac{\Delta t_{FD}}{\Delta t_M} S_{atm}(t_{FD}) \mu(t_M)}{\sum_{i=1}^{\frac{\Delta t_{FD}}{\Delta t_M}} \mu(t_{M_i})} && \text{for } \mu(t_M) > 0.001 \\ S_{atm}(t_M) &= 0 && \text{for } \mu(t_M) \leq 0.001 \end{aligned} \quad (2.1007)$$

where Δt_{FD} is the time step of the forcing data (3 hours \times 3600 seconds hour $^{-1}$ = 10800 seconds), Δt_M is the model time step (seconds), $S_{atm}(t_{FD})$ is the three-hourly solar radiation from the forcing data (W m^{-2}), and $\mu(t_M)$ is the cosine of the solar zenith angle at model time step t_M (section 2.3.3). The term in the denominator of equation (1) is the sum of the cosine of the solar zenith angle for each model time step falling within the three hour period. For numerical purposes, $\mu(t_{M_i}) \geq 0.001$.

The total incident solar radiation S_{atm} at the model time step t_M is then split into near-infrared and visible radiation and partitioned into direct and diffuse according to factors derived from one year's worth of hourly CAM output from CAM version cam3_5_55 as

$$S_{atm} \downarrow_{vis}^\mu = R_{vis} (\alpha S_{atm}) \quad (2.1008)$$

$$S_{atm} \downarrow_{nir}^\mu = R_{nir} [(1 - \alpha) S_{atm}] \quad (2.1009)$$

$$S_{atm} \downarrow_{vis} = (1 - R_{vis}) (\alpha S_{atm}) \quad (2.1010)$$

$$S_{atm} \downarrow_{nir} = (1 - R_{nir}) [(1 - \alpha) S_{atm}] . \quad (2.1011)$$

where α , the ratio of visible to total incident solar radiation, is assumed to be

$$\alpha = \frac{S_{atm} \downarrow_{vis}^\mu + S_{atm} \downarrow_{vis}}{S_{atm}} = 0.5. \quad (2.1012)$$

The ratio of direct to total incident radiation in the visible R_{vis} is

$$R_{vis} = a_0 + a_1 \times \alpha S_{atm} + a_2 \times (\alpha S_{atm})^2 + a_3 \times (\alpha S_{atm})^3 \quad 0.01 \leq R_{vis} \leq 0.99 \quad (2.1013)$$

and in the near-infrared R_{nir} is

$$R_{nir} = b_0 + b_1 \times (1 - \alpha) S_{atm} + b_2 \times [(1 - \alpha) S_{atm}]^2 + b_3 \times [(1 - \alpha) S_{atm}]^3 \quad 0.01 \leq R_{nir} \leq 0.99 \quad (2.1014)$$

where $a_0 = 0.17639$, $a_1 = 0.00380$, $a_2 = -9.0039 \times 10^{-6}$, $a_3 = 8.1351 \times 10^{-9}$ and $b_0 = 0.29548$, $b_1 = 0.00504$, $b_2 = -1.4957 \times 10^{-5}$, $b_3 = 1.4881 \times 10^{-8}$ are coefficients from polynomial fits to the CAM data.

The additional atmospheric forcing variables required by Table 2.4 are derived as follows. The atmospheric reference height z'_{atm} (m) is set to 30 m. The directional wind components are derived as $u_{atm} = v_{atm} = W_{atm}/\sqrt{2}$. The potential temperature $\overline{\theta}_{atm}$ (K) is set to the atmospheric temperature T_{atm} . The atmospheric longwave radiation $L_{atm} \downarrow$ (W m^{-2}) is derived from the atmospheric vapor pressure e_{atm} and temperature T_{atm} ([Idso 1981](#)) as

$$L_{atm} \downarrow = \left[0.70 + 5.95 \times 10^{-5} \times 0.01 e_{atm} \exp\left(\frac{1500}{T_{atm}}\right) \right] \sigma T_{atm}^4 \quad (2.1015)$$

where

$$e_{atm} = \frac{P_{atm}q_{atm}}{0.622 + 0.378q_{atm}} \quad (2.1016)$$

and σ is the Stefan-Boltzmann constant ($\text{W m}^{-2} \text{K}^4$) (Table 2.7). The fraction of precipitation P (mm s^{-1}) falling as rain and/or snow is

$$q_{rain} = P(f_P), \quad (2.1017)$$

$$q_{snow} = P(1 - f_P) \quad (2.1018)$$

where

$$f_P = 0 < 0.5(T_{atm} - T_f) < 1. \quad (2.1019)$$

The aerosol deposition rates D_{sp} (14 rates as described in Table 2.4) are provided by a time-varying, globally-gridded aerosol deposition file developed by *Lamarque et al. (2010)*.

If the user wishes to provide atmospheric forcing data from another source, the data format outlined above will need to be followed with the following exceptions. The data atmosphere model will accept a user-supplied relative humidity RH (%) and derive specific humidity q_{atm} (kg kg^{-1}) from

$$q_{atm} = \frac{0.622e_{atm}}{P_{atm} - 0.378e_{atm}} \quad (2.1020)$$

where the atmospheric vapor pressure e_{atm} (Pa) is derived from the water ($T_{atm} > T_f$) or ice ($T_{atm} \leq T_f$) saturation vapor pressure $e_{sat}^{T_{atm}}$ as $e_{atm} = \frac{RH}{100}e_{sat}^{T_{atm}}$ where T_f is the freezing temperature of water (K) (Table 2.7), and P_{atm} is the pressure at height z_{atm} (Pa). The data atmosphere model will also accept a user-supplied dew point temperature T_{dew} (K) and derive specific humidity q_{atm} from

$$q_{atm} = \frac{0.622e_{sat}^{T_{dew}}}{P_{atm} - 0.378e_{sat}^{T_{dew}}}. \quad (2.1021)$$

Here, e_{sat}^T , the saturation vapor pressure as a function of temperature, is derived from *Lowe's (1977)* polynomials. If not provided by the user, the atmospheric pressure P_{atm} (Pa) is set equal to the standard atmospheric pressure $P_{std} = 101325$ Pa, and surface pressure P_{srf} (Pa) is set equal to P_{atm} .

The user may provide the total direct and diffuse solar radiation, $S_{atm} \downarrow^\mu$ and $S_{atm} \downarrow$. These will be time-interpolated using the procedure described above and then each term equally apportioned into the visible and near-infrared wavebands (e.g., $S_{atm} \downarrow_{vis}^\mu = 0.5S_{atm} \downarrow^\mu$, $S_{atm} \downarrow_{nir}^\mu = 0.5S_{atm} \downarrow^\mu$).

2.32.1 Anomaly Forcing

The ‘Anomaly Forcing’ atmospheric forcing mode provides a means to drive CLM with projections of future climate conditions without the need for large, high-frequency datasets. From an existing climate simulation spanning both the historical and future time periods, a set of anomalies are created by removing a climatological seasonal cycle based on the end of the historical period from each year of the future time period of the simulation. These anomalies can then be applied to a repeating high-frequency forcing dataset of finite duration (e.g. 10 years). State and flux forcing variables are adjusted using additive and multiplicative anomalies, respectively:

$$\begin{aligned} S' &= S + k_{anomaly} && \text{state variable} \\ F' &= f \times k_{anomaly} && \text{flux variable} \end{aligned} \quad (2.1022)$$

where S' is the adjusted atmospheric state variable, S is the state variable from the high-frequency reference atmospheric forcing dataset, and $k_{anomaly}$ is an additive anomaly. Similarly, F' is the adjusted atmospheric flux variable, F is the flux variable from the high-frequency reference atmospheric forcing dataset, and $k_{anomaly}$ is a multiplicative anomaly. State variables are temperature T_{atm} , pressure P_{atm} , humidity q_{atm} , and wind W_{atm} . Flux variables are precipitation P , atmospheric shortwave radiation $S_{atm} \downarrow$, and atmospheric longwave radiation $L_{atm} \downarrow$.

2.33 References

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CHAPTER 3

INDICES AND TABLES

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