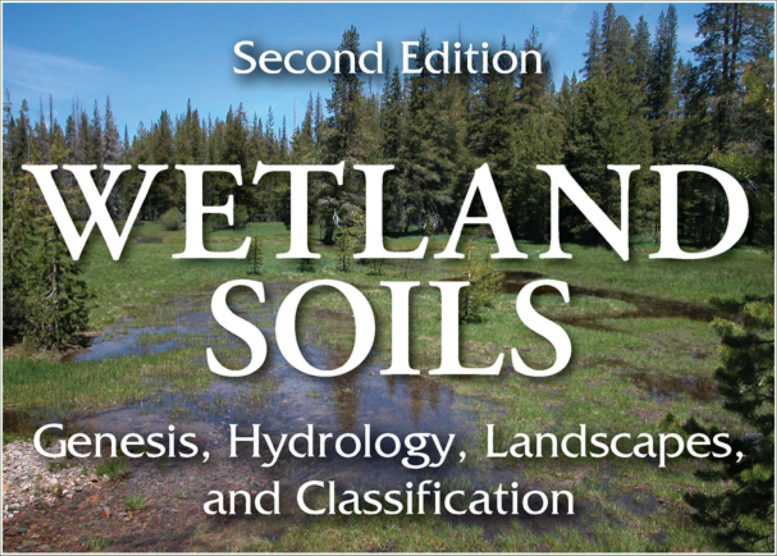


Second Edition

WETLAND SOILS

Genesis, Hydrology, Landscapes, and Classification

Edited by

Michael J. Vepraskas • Christopher B. Craft

CRC Press
Taylor & Francis Group

Second Edition

WETLAND SOILS

Genesis, Hydrology, Landscapes, and Classification

Second Edition

WETLAND SOILS

Genesis, Hydrology, Landscapes, and Classification

Edited by

Michael J. Vepraskas • Christopher B. Craft

CRC Press
Taylor & Francis Group
Boca Raton London New York

CRC Press is an imprint of the
Taylor & Francis Group, an **informa** business

CRC Press
Taylor & Francis Group
6000 Broken Sound Parkway NW, Suite 300
Boca Raton, FL 33487-2742

© 2016 by Taylor & Francis Group, LLC
CRC Press is an imprint of Taylor & Francis Group, an Informa business

No claim to original U.S. Government works
Version Date: 20151002

International Standard Book Number-13: 978-1-4398-9800-0 (eBook - PDF)

This book contains information obtained from authentic and highly regarded sources. Reasonable efforts have been made to publish reliable data and information, but the author and publisher cannot assume responsibility for the validity of all materials or the consequences of their use. The authors and publishers have attempted to trace the copyright holders of all material reproduced in this publication and apologize to copyright holders if permission to publish in this form has not been obtained. If any copyright material has not been acknowledged please write and let us know so we may rectify in any future reprint.

Except as permitted under U.S. Copyright Law, no part of this book may be reprinted, reproduced, transmitted, or utilized in any form by any electronic, mechanical, or other means, now known or hereafter invented, including photocopying, microfilming, and recording, or in any information storage or retrieval system, without written permission from the publishers.

For permission to photocopy or use material electronically from this work, please access www.copyright.com (http://www.copyright.com/) or contact the Copyright Clearance Center, Inc. (CCC), 222 Rosewood Drive, Danvers, MA 01923, 978-750-8400. CCC is a not-for-profit organization that provides licenses and registration for a variety of users. For organizations that have been granted a photocopy license by the CCC, a separate system of payment has been arranged.

Trademark Notice: Product or corporate names may be trademarks or registered trademarks, and are used only for identification and explanation without intent to infringe.

Visit the Taylor & Francis Web site at
http://www.taylorandfrancis.com

and the CRC Press Web site at
http://www.crcpress.com

Contents

Preface

The second edition of *Wetland Soils* updates the information in the first edition without changing the focus from a field orientation. While many chapters included here are found in the first edition, all have been revised. New authors were brought on board when original authors either retired or changed their focus. Three new chapters have been added to enhance information on wetland functions and restoration.

Anyone dealing with wetlands should understand the properties and functions of the soils found in and around wetlands. The ability to identify wetland soils is at the core of wetland delineation. Wetland restoration revolves around techniques that are designed to restore the physical properties and chemical reactions that occur in these soils. These chemical processes cause wetland soils to become anaerobic, supporting microbial activity not found in terrestrial soils, and requiring special adaptations of plants if they are to survive in wetland environments.

The contributors describe a diverse range of soils that occur in and around wetlands throughout North America. These wetlands are widely recognized as consisting three main components: hydric soils, hydrophytic vegetation, and wetland hydrology. While most wetlands could be identified and their functions understood if the site's hydrology were known, an individual wetland's hydrology is far too dynamic for field workers to understand fully without long-term monitoring studies. Some morphological aspects of hydric soils, however, can be used to evaluate a site's hydrology. This book explains how soil morphology can be used as a field tool to evaluate soil hydrology and soil biogeochemical processes. A recurring theme in this book is that hydric soils are components of a landscape, whose soils have been altered by hydrologic and biogeochemical processes.

In keeping with the first edition, the book is organized into three sections. Section I examines the basic concepts, processes, and properties that pertain to virtually any hydric soil. We recognize that most users of this book will not be soil scientists, so, Chapter 1 is a general overview that introduces important terms and concepts. Chapter 2 explains the historic development of the concept of hydric soil while the following chapters examine soil hydrology, chemistry, biology, soil organic matter, and the development and use of the hydric soil field indicators.

Section II of the book is devoted to the soils in specific kinds of wetlands. Here, we have classified wetlands based on the landscape or geomorphic position, when possible. This section does not include all wetlands, but rather focuses on those that we felt had unique aspects that needed further elaboration. Section III is devoted to wetland functional analysis and the restoration of freshwater and tidal wetlands.

The terminology used throughout the book is that used by soil scientists. Soils are described and classified according to the conventions of the USDA's Natural Resources Conservation Service. Common wetland terms such as *fen*, *peatland*, or *pocosin* are used only to illustrate a particular concept.

As with the first edition, we intend this to be a comprehensive book on hydric soils that could be used as a text in college courses and as a reference for practicing professionals.

Our hope is that this book will improve communication among soil scientists, hydrologists, and biologists, and will prepare individuals to work with real wetlands in the field.

M. J. Vepraskas and C. B. Craft
Editors

Editors

Michael J. Vepraskas is a William Neal Reynolds Distinguished Professor of Soil Science at North Carolina State University, where he is also head of the Soil Science Department. Currently, he specializes in wetlands with a focus on hydric soil formation and identification. His class on *Wetland Soils* is taught both on campus and online. In addition, he trains wetland consultants and regulators in identifying hydric soils through short courses that have been taught throughout the United States. His research has developed techniques for identifying wetlands that are used nationwide to enforce both state and federal regulations. He is a member of the National Technical Committee for Hydric Soils, which determines the policies for how hydric soils are defined and identified by law throughout the United States. Mike holds degrees from the University of Wisconsin in Madison (BS and MS) and Texas A&M University (PhD). Professor Vepraskas has been an editor for the *Soil Science Society of America Journal, Wetlands, and Geoderma*, and is a fellow of Soil Science Society of America, American Society of Agronomy, Society of Wetland Scientists, and the American Association for the Advancement of Science. He is also an adjunct professor of soil science at Virginia Tech University and an adjunct professor of geology at the University of Tennessee.

Christopher B. Craft is the Janet Duey Professor of rural land policy in the School of Public and Environmental Affairs at Indiana University, Bloomington, where he teaches courses in Wetlands Ecology, Restoration Ecology, Applied Ecology, and Environmental Science. He earned degrees from the University of North Carolina-Asheville (BA, biology), the University of Tennessee (MS, ecology), and North Carolina State University (PhD, soil science). Professor Craft has over 30 years of experience working in inland freshwater and coastal wetlands. His research interests include wetland restoration, nutrient enrichment and eutrophication, carbon sequestration, and effects of climate change; his research projects span the Eastern, Midwest, and Western United States, Europe, and China. Since 2010, Professor Craft has been a visiting professor with Senior International Scientists of the Chinese Academy of Sciences. Chris has been a founding member (since 2000) of the Georgia Coastal Ecosystems Long Term Ecological Research program (GCE LTER), and has been serving on its Executive Committee since 2012. He served as president of the Society of Wetland Scientists (SWS), an international organization that promotes sound wetland science, education, and management, from 2008 to 2009. In 2012, Professor Craft received the National Wetlands Award for Science Research sponsored by the Environmental Law Institute, Washington, DC.

Contributors

Anne E. Altor
Precision Science Editing
Carmel, Indiana

James L. Arndt
URS Corporation
Bloomington, Minnesota

Jay C. Bell
Department of Soil, Water, and Climate
University of Minnesota
St. Paul, Minnesota

Jacob F. Berkowitz
Environmental Laboratory
U.S. Army Corps of Engineers
Vicksburg, Mississippi

Janis L. Boettinger
Department of Plant, Soils, and Climate
Utah State University
Logan, Utah

Scott D. Bridgham
Institute of Ecology and Evolution
University of Oregon
Eugene, Oregon

Gregory L. Bruland
Biology and Natural Resources Department
Principia College
Elsah, Illinois

Christopher B. Craft
School of Public and Environmental
Affairs
Indiana University
Bloomington, Illinois

Patrick J. Drohan
Department of Ecosystem Science
and Management
Penn State University
University Park, Pennsylvania

Ryan E. Emanuel
Department of Forestry and
Environmental Resources
North Carolina State University
Raleigh, North Carolina

Stephen P. Faulkner
Leetown Science Center
U.S. Geological Survey
Kearneysville, West Virginia

M. Siobhan Fennessy
Biology Department
Kenyon College
Gambier, Ohio

Matthew F. Hanchey
Nicholas School of the Environment
Duke University
Durham, North Carolina

Willie G. Harris Jr.
Department of Soil and Water Science
University of Florida
Gainesville, Florida

Ellen R. Herbert
School of Public and Environmental
Affairs
Indiana University
Bloomington, Indiana

G. Wade Hurt
Department of Soil and Water Science
University of Florida
Gainesville, Florida

Jason K. Keller
Schmid College of Science
and Technology
Chapman University
Orange, California

I.T. Kenney
Department of Agronomy
Kansas State University
Manhattan, Kansas

Randy Kolka
USDA Forest Service
Northern Research Station
Grand Rapids, Minnesota

David L. Lindbo
Department of Soil Science
North Carolina State University
Raleigh, North Carolina

John M. Marton
CDM Smith
Indianapolis, Indiana

Cassandra A. Medvedeff
Schmid College of Science
and Technology
Chapman University
Orange, California

Aaron J. Miller
USDA-Natural Resources Conservation
Service
Santa Fe Soil Survey Office
Santa Fe, New Mexico

Brian A. Needelman
Department of Environmental Science
and Technology
University of Maryland
College Park, Maryland

Christopher V. Noble
Environmental Laboratory
U.S. Army Corps of Engineers
Vicksburg, Mississippi

Chien-Lu Ping
School of Natural Resources and Extension
University of Alaska Fairbanks
Palmer, Alaska

Matthew Polizzotto
Department of Soil Science
North Carolina State University
Raleigh, North Carolina

Deann R. Presley
Department of Agronomy
Kansas State University
Manhattan, Kansas

Martin C. Rabenhorst
Department of Environmental Science
and Technology
University of Maryland
College Park, Maryland

Curtis J. Richardson
Nicholas School of the Environment
Duke University
Durham, North Carolina

Jimmie L. Richardson
Department of Soil Science
North Dakota State University
Fargo, North Dakota

Steven W. Sprecher
USDA-Natural Resources Conservation
Service
Geneseo, New York

Ariana E. Sutton-Grier
Cooperative Institute for Climate and
Satellites, Earth System Science
University of Maryland and NOAA
Silver Spring, Maryland

James A. Thompson
Division of Plant and Soil Sciences
West Virginia University
Morgantown, West Virginia

Lenore M. Vasilas
USDA-Natural Resources Conservation
Service
Soil Survey Division
Bethesda, Maryland

Karen L. Vaughan
Ecosystem Science and Management Department
University of Wyoming
Laramie, Wyoming

Michael J. Vepraskas
Department of Soil Science
North Carolina State University
Raleigh, North Carolina

Denise Wardrop
Department of Geography
Penn State University
University Park, Pennsylvania

Frank C. Watts
AAA Soil Consultants
Hilliard, Florida

Section I

Basic Principles of Hydric Soils

1

Basic Concepts of Soil Science

Deann R. Presley, Steven W. Sprecher, and I.T. Kenney

CONTENTS

Introduction

This chapter provides an introduction to soil description in the field, soil classification, and soil survey. Understanding and interpreting soils is an iterative process that begins with describing a single soil profile, then a consideration of the landscape position, hydrology, and soil formation processes active upon the soil. Ultimately, this process ends with an assessment of the soil's potential use.

The terminology and approach used are those of the Soil Survey Staff of the U.S. Department of Agriculture Natural Resources Conservation Service (USDA-NRCS), the federal agency with primary responsibilities for defining and cataloging soils in the United States. The topics covered include the information necessary to complete the soils portion of wetland delineation forms and some common soil science terminology that experience has shown may be misunderstood by wetland scientists who have had no formal training in soil science. It is also important to learn and use these standard terms and classification so that a trained professional would be able to understand the profile without having to see the soil personally, perhaps several years after the soil was described.

The various disciplines that study soils define "soil" according to how they use it. Civil engineers emphasize physical properties; geologists emphasize degree of weathering; and agriculturalists focus on the properties of soil as a growth medium. "Pedology" is the branch of soil science that studies the components and formation of soils, assigning them taxonomic status, and mapping and explaining soil distributions across the landscape. It provides the perspective from which the USDA Soil Survey Program regards soils and is also the perspective of this book. A pedologic definition of soil is

> soil (i) The unconsolidated mineral or organic material on the immediate surface of the earth that serves as a natural medium for the growth of land plants. (ii) The unconsolidated mineral or organic matter on the surface of the earth that has been subjected to and shows effects of genetic and environmental factors of: climate (including water and temperature effects), and macro- and microorganisms, conditioned by relief, acting on parent material over a period of time. A product-soil differs from the material from which it is derived in many physical, chemical, biological, and morphological properties and characteristics.
>
> **Soil Science Society of America 2008.**

Here, soil is seen to have a natural organization and to be biologically active. This inherent organization results from climatic and biological forces altering the properties of the materials of the earth's surface. Because these soil-forming forces exert progressively less influence with depth, they result in more or less horizontal layers that are termed "soil horizons" (Figure 1.1). Individual kinds of soil are distinguished by their specific sequence of horizons, or "soil profile." The characteristics and vertical sequences of these soil horizons vary in natural patterns across the landscape.

Organic Soils and Mineral Soils

There are two major categories of soil parent materials, organic and mineral. Organic soils form from plant debris. These soils are found in wetlands because plant debris decomposes less rapidly in very wet settings, due to low oxygen conditions, also referred to as "anaerobic." Organic soils are very black, porous, and light in weight, and are often referred to as "peats" or "mucks."

Mineral soils, on the other hand, form from rocks in place or from material transported by wind, water, landslide, or ice. Consequently, mineral soil materials consist of different amounts of sand, silt, and clay (described later in this chapter), and constitute the majority of the soils in the world. They occur both within and outside of wetlands.

In practice, mineral and organic soils are separated on the basis of organic carbon levels. The threshold carbon contents separating organic and mineral soils are shown in Figure 1.2. Organic matter concentrations above these levels dominate the physical and chemical properties of the soil. It is very difficult to estimate organic carbon content in the field without training using samples of known carbon concentration. In general, if the soil feels gritty or sticky, or resists compression, it is mineral material; if the soil material feels slippery or greasy when rubbed, has almost no internal strength, and stains the fingers, it may be organic. If the material is an organic soil material, a further division should be made, and is described in the following section.

Soil Horizons

As previously noted, soils are separated largely on the basis of the types of horizons they have and horizon properties. Horizons, in turn, are differentiated from each other by

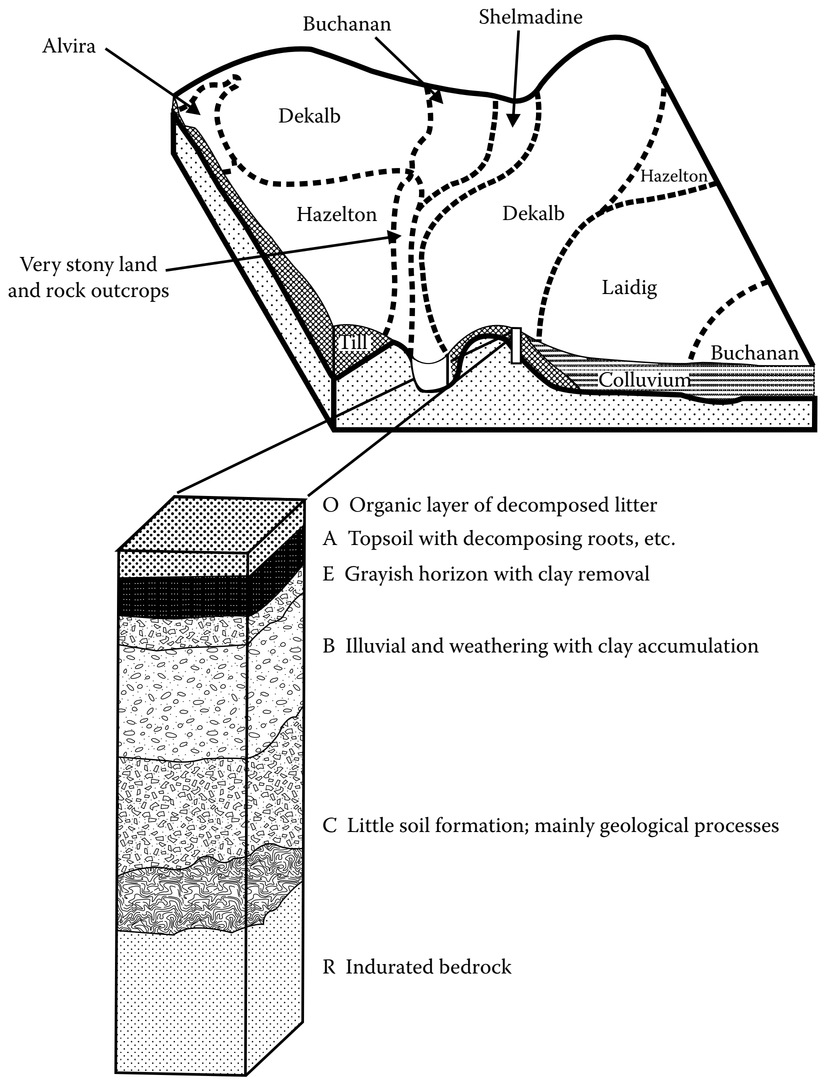


FIGURE 1.1
Hypothetical soil profile with master horizons (O, A, E, B, C, and R horizons) and the surrounding landscape, including other mapped soils on the landscape (dashed lines). (Adapted from Lipscomb, G. H. 1992. *Soil Survey of Monroe County, Pennsylvania.* USDA-SCS in cooperation with the Pennsylvania State University and Pennsylvania Department of Environmental Resources, U.S. Government Printing Office, Washington, DC.)

differences in organic carbon content, morphology (color, texture, structure, etc.), mineralogy, and chemistry (pH, redox regime, etc.). Most people are aware that mineral soils in general have a dark, friable topsoil and lighter colored, firmer subsoil. Below the subsoil is geologic material that has not yet weathered into soil; this may be alluvium, decomposed rock, unweathered bedrock, or other materials.

In very general terms, pedologists call the topsoil in mineral soils the “A horizon,” the subsoil the “B horizon,” the underlying parent material the “C horizon,” and unweathered rock, the “R horizon” (Figure 1.1). Pedologists also recognize a light-colored “E horizon” that may be present between the A and B horizons. Organic soils contain organic horizons

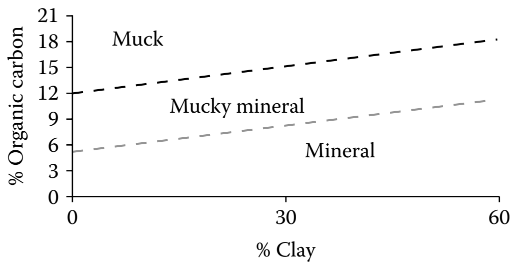


FIGURE 1.2
Levels of clay and organic carbon that define distinctions between organic and mineral soil materials (bold font). An uncommon but important subset of mineral materials is "mucky mineral" soil materials (carbon and clay contents between the dashed lines). (From United States Department of Agriculture, Natural Resources Conservation Service. 2010. In L. M. Vasilas, G.W. Hurt, and C.V. Noble (Eds.) *Field Indicators of Hydric Soils in the United States, Version 7.0*. USDA, NRCS, in cooperation with the National Technical Committee for Hydric Soils. Available at: ftp://ftp-fc.sc.egov.usda.gov/NSSC/Hydric_Soils/FieldIndicators_v7.pdf)

("O horizons"). Each kind of master horizon (A, B, C, E, and O horizon) may be subdivided into different subhorizons. Approximately 22,000 named soils in the United States are differentiated from each other on the basis of the presence and sequence of these different subhorizons, as well as external factors such as climate, hydrologic regime, and parent material. Pedologists study the earth's surface to a depth of about 2 m; parent material differences at greater depths usually are not considered.

O Horizons and Organic Soils

An O horizon is a rather common feature of wetland soils, because of slow decomposition of the abundant plant materials that often grow in the wetland environment. There are three subscripts that are exclusively used with O horizons. To choose the most appropriate subscript, rub a moist sample of soil between the fingers 10 times, and then observe the percent of visible plant fibers with a hand lens. This is called the rubbed state. Live roots are not included in this determination (Table 1.1).

The USDA-NRCS currently recognizes three classes of organic matter for field description of soil horizons: sapric, hemic, and fibric materials. Differentiating criteria are based

TABLE 1.1

Percent Volume Fiber Content of Sapric, Hemic, and Fibric Organic Soil Horizons

Horizon Descriptor	Horizon Symbol	Proportion of Fibers Visible with a Hand Lens: Unrubbed	Rubbed
Sapric	Oa	<1/3	<1/6
Hemic	Oe	1/3–2/3	1/6–2/5
Fibric	Oi	>2/3	>2/5

Source: Adapted from Soil Survey Staff. 1975. *Soil Taxonomy: A Basic System of Soil Classification for Making and Interpreting Soil Surveys. USDA-SCS Agricultural Handbook 436*. U.S. Government Printing Office, Washington, DC.

on the percent of visible plant fibers observable with a hand lens (i) in an unrubbed state and (ii) after rubbing between the thumb and fingers 10 times (Table 1.1). "Sapric," "hemic," and "fibric" roughly correspond to the older terms "muck," "mucky peat," and "peat," respectively. Complete details on identifying sapric, hemic, and fibric materials are given in Chapter 8.

Soil Descriptions for Wetland Delineation Forms

The level of detail employed in describing and differentiating soil horizons varies with the purpose of the soil study. Wetland studies and determinations focus on depths of evidence of anaerobiosis and iron reduction. Consequently, changes with depth in soil color are more important than subtle changes in soil texture and structure for hydric soil determinations. *Field Indicators of Hydric Soils of the United States* (NTCHS 2010; hereafter "The NTCHS Field Indicators") refers to soil "layers" rather than to horizons because of the focus on shallow hydrologic regimes. Multiple pedogenic horizons often may be lumped into fewer colorimetric layers for our purposes. Hence, the absence of data cells soliciting information about soil structure, for example, in federal wetland determination data forms.

Hydric soil data form solicit information about horizon depths, color, redoximorphic features (formerly called mottles), and an estimate of texture (see Chapter 8). Other features, too, should be described if pertinent to the investigation. Formal procedures for describing soils can be found in the *Soil Survey Manual* (Soil Survey Division Staff 1993) and the *Field Book for Describing and Sampling Soils* (Schoeneberger et al. 2012).

Depth from the soil surface is critical in making hydric soil decisions based on the NTCHS Field Indicators, which defines the soil surface differently for different land resource regions around the country (see Chapter 9). The depth of the top and bottom of each horizon is recorded when describing soils; the top of the first horizon is the soil surface. Subsequent horizons are distinguished from those above by change in soil color, texture, or structure, or by changes in the presence or absence of redoximorphic features.

Soil Colors

The most obvious feature of a soil body or profile is its color. Extremely important site characteristics, such as a hydrologic regime, mineral weathering, and water content, can be inferred from soil color. Organic matter darkens the soil and is typically associated with surface layers, usually masking all other coloring agents. Well-drained soils often have uniform and bright colors. Iron is the primary coloring agent in the subsoil. The orange brown colors associated with well-drained soils are the result of iron oxide stains that coat individual particles. Soils with a fluctuating water table usually have a mottled, or spotted, pattern of gray, yellow (Y), and/or orange colors (described in a later section of this chapter). Soils with a high water table for a significant portion of the year have very gray background or matrix colors. The word gley is used for gray colors resulting from water logging and iron reduction.

The discipline of soil science in the United States uses the Munsell color system to quantify color in a standard, reproducible manner. The *Munsell® Soil Color Charts* (Munsell Color 2013) are contained in a 15 × 20-cm six-ring binder of 12 pages, or charts. Each chart consists

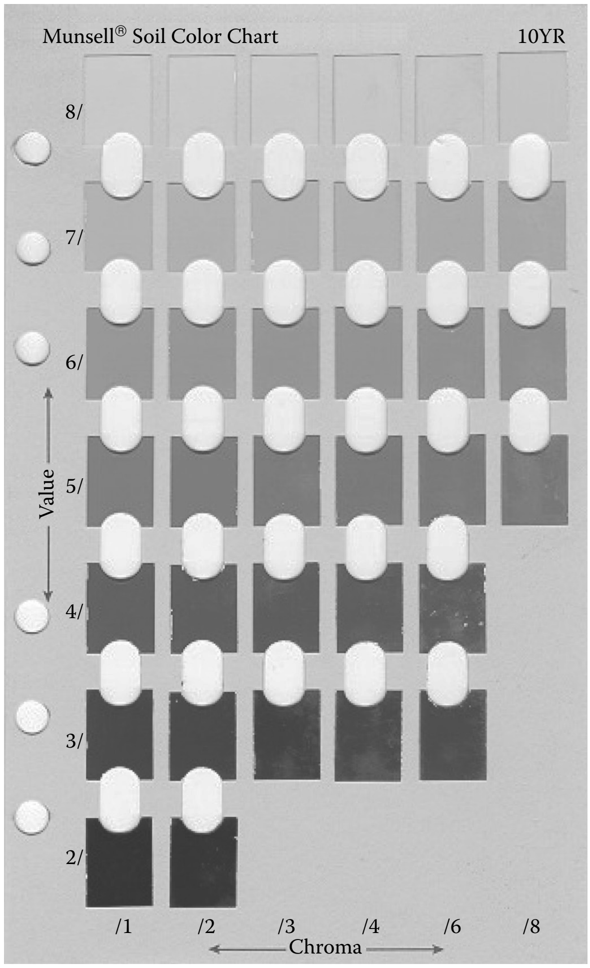


FIGURE 1.3
(**See color insert**.) The 10YR page from the *Munsell® Soil Color Chart*.

of 29–42 color chips. The Munsell system notes three aspects of color, in the sequence "Hue Value/Chroma," for example, 10YR 4/2 (Figure 1.3). All the chips on an individual chart have the same hue. Within a particular hue—that is, on any one color chart—values are arrayed in rows and chromas in columns. Hue can be thought of as the spectral color (quality of pigmentation), value can be thought of as the lightness or darkness, and chroma can be thought of as the richness of pigmentation (pale to bright).

Specifically, hue describes how much red (R), Y, green (G), blue (B), or purple (P) is in a color. The degree of redness or yellowness, etc., is quantified with a number preceding the letter, for example, 2.5Y. Most soil hues are combinations of R and Y, which we perceive as shades of brown. These differences in hue are organized in the Munsell color charts from reddest (10R) to yellowest (5Y), with the chips of each hue occupying one page of the charts. The sequence of charts, from reddest to yellowest, is as follows:

10R 2.5YR 5YR 7.5YR 10YR 2.5Y 5Y

Reddest Red - yellow mixes Yellowest

Many soils in the United States have 10YR hues; so, start with that chart unless most of the soils in your local area have different hues. Subsoils containing minerals with reduced iron (Fe^{2+}) may be yellower or greener than hue 5Y. Such colors are represented on the color charts for gley, or the "gley pages" (Figure 1.4). These have neutral hue (N) or hues of Y, G, B, or P. Soil layers with colors found on the gley charts are generally saturated with water for very long periods of time and may be found in wetlands (USDA-NRCS 2010).

Value denotes darkness and lightness, or simply the amount of light reflected by the soil or a color chip. A-horizon colors usually have a low value (very dark to black) because of staining by organic matter. Colors of hydric soil field indicators frequently need to be determined below the zone of organic staining where values are higher than 3 or 4; the exceptions are when a hydric soil feature is made up of organic matter, or when thick, dark A-horizon materials continue down the soil profile for several decimeters (Chapter 8).

Chroma quantifies the richness of pigmentation or concentration of hue. High-chroma colors are richly pigmented; low-chroma colors have little pigmentation and are dull and grayish. Subsoil layers that are waterlogged and chemically reduced much of the year

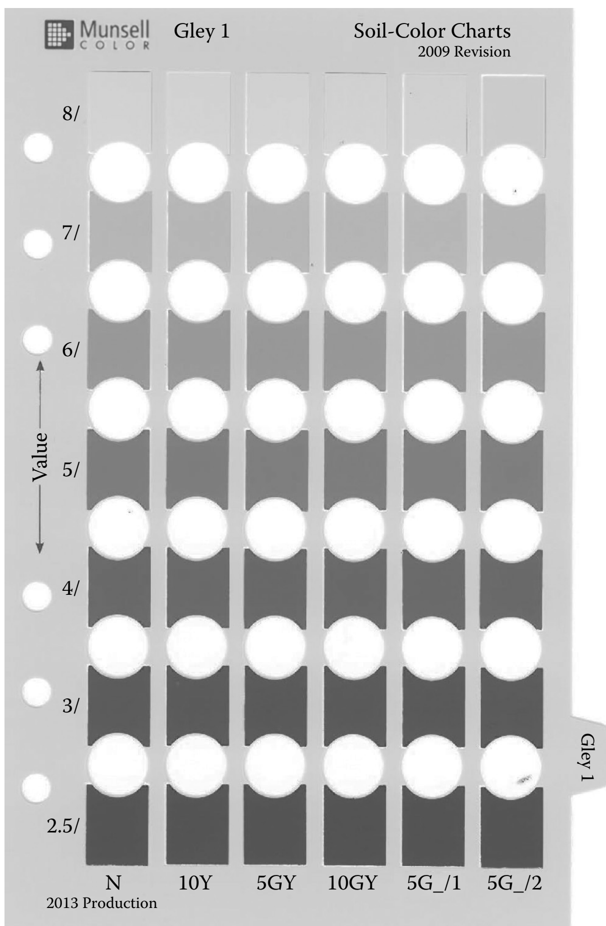


FIGURE 1.4
(**See color insert.**) Example of the gley page from the *Munsell® Soil Color Chart*.

have much of their iron-based pigmentation “washed out” of them and have low-chroma colors.

Soil colors seldom match any Munsell color chip perfectly. Standard NRCS procedures require that Munsell colors be read to the nearest chip and not be interpolated between chips. Recent NRCS guidance for hydric soil determination, however, requires that colors be noted as equal to, greater than, or less than critical color chips (USDA-NRCS 2010; see also Chapter 8). For hydric soil determinations, colors should not be extrapolated beyond the range of chips in the color book.

Soil samples should be read under standard conditions because soil colors vary with differences in light quality, moisture content, and sample condition. Color charts are designed to be read in full, mid-day sunlight, because soils appear redder when the sun is lower in the sky. Stand with the sun at your back so that the sunlight strikes the soil sample and color chips at a right angle. And remember to remove your sunglasses when reading soil colors.

Describe soils on the basis of moist colors. To bring a soil specimen to the moist state, slowly spray water onto the sample until it no longer changes color. The soil is too wet if it glistens; allow it to dry until its surface is dull. Break the soil specimen open gently, and read the color off the otherwise-undisturbed, broken face. Both the inside and outside of natural soil aggregates can be read in this way.

Matrix and Special Features

The predominant color of a soil horizon is known as its matrix color, that is, the color that occupies more than half of the volume of the horizon. If a horizon has several colors and none of it occupies 50% of the volume, describe the various colors and report the percent volume for each. Often, soil aggregates have different exterior and interior colors; these, too, are noted separately.

Mottles are small areas that differ in color from the soil matrix. Mottles that result from waterlogging and chemical reduction are called redoximorphic features. These features are listed as part of the field indicators for hydric soils and should be described carefully when filling out wetland data forms (Vepraskas 2004). Chemical reduction is not the only source of color differences within the soil. Other causes of color differences within a layer include recently sloughed root material (often reddish), root decomposition (very dark gray to black), decomposition of pebbles or rocks (usually an abrupt, strong contrast with the surrounding matrix), and carbonate accumulation (white).

Standard USDA-NRCS soil sampling protocols require a description of mottle color, abundance, size, contrast, and location (Schoeneberger et al. 2012). Colors of redoximorphic features should be described with the standard Munsell notation. Federal hydric soil determination forms do not solicit size or contrast of redoximorphic features, and replace abundance with numeric percentages of volume occupied. Percent volume should be determined using diagrams for estimating proportions of mottles; these usually accompany commercial soil color books and can also be found in reference materials, such as the USDA-NRCS field manual by Schoeneberger et al. (2012). Most people overestimate the abundance of mottles without the use of some aid.

For hydric soils work, Munsell color difference between redoximorphic features and the surrounding matrix is critical and is evaluated quantitatively. The distinction between faint and distinct color contrast is important enough in the NTCHS Field Indicators that the difference can determine whether a soil is hydric or not. Note, however, that only Munsell colors are solicited on most federal hydric soil determination forms, and not differences in color or contrast classes. The legal implications of many hydric soil determinations should motivate us to record field data as objectively as possible and to make hydric soil decisions *after* a soil pit is described. Leaving contrast classes off the field form imposes one extra step between description and interpretation.

Federal wetland determination data forms must also denote where redoximorphic features occur, such as if they are located within the matrix or if they occur on pore linings; pore linings include both root channels and ped faces or fracture planes (see Chapter 7 for further details).

The following is an example for determining the degree of contrast of redoximorphic features (RMF). If the matrix color is 10YR 4/4, then, an RMF of color 10YR 6/4 (Δhue = 0, Δvalue = 2, and Δchroma = 0) would have a faint contrast, while an RMF of color 7.5YR 6/4 (Δhue = 1, Δvalue = 2, and Δchroma = 0) would have a distinct contrast, and an RMF of color 5YR 6/6 (Δhue = 2, Δvalue = 2, and Δchroma = 2) would have a prominent contrast (Table 1.2).

Soil Texture

The relative percentage of sand, silt, and clay in a soil sample is referred to as "soil texture." Soil texture is usually recorded for each soil horizon. Unfortunately, the terminology of soil texture is confusing because some of the same terms are used to describe both (1) individual soil particles and (2) mixtures of particles (e.g., sand, silt, and clay) and (3) a class of soil minerals (clay).

Individual mineral particles range in size from boulders to microscopic clay crystals. Soil textures for USDA Soil Survey are determined on the basis of particles having diameters of 2 mm and smaller. The USDA soil texture system identifies three classes of particles: sand, silt, and clay (Table 1.3). A fourth class, coarse fragments, is also recognized (i.e., gravels >2 mm, rocks, etc.), but coarse fragments are disregarded when determining the USDA texture of a soil.

Sand particles feel at least slightly gritty when rubbed between the fingers and often are audible when rubbed near your ear. Silt materials feel like flour when rubbed. Most clays

TABLE 1.2

Abundance, Size, and Contrast of Mottles

Mottle Abundance[a]		Mottle Size[a]	
Few	<2%	Fine	<5 mm
Common	2%–20%	Medium	5–15 mm
Many	>20%	Coarse	>15 mm
Mottle Contrast[b]			
	Hues on the Same Chart (e.g., Both Colors 10YR)	Hue Difference on a Chart (e.g., 10YR vs. 7.5YR)	Hue Difference on Two Charts or More (e.g., 10YR vs. 5YR)
Faint	≤2 units of value, and ≤1 unit of chroma	≤1 unit of value and ≤1 unit of chroma	Hue differences of two or more charts are distinct or prominent
Distinct	Between faint and prominent	Between faint and prominent	0 to <2 units of chroma and/or value
Prominent	At least 4 units in value and/or chroma	At least 3 units in value and/or chroma	At least 2 units in value and/or chroma

[a] From Soil Survey Division Staff (1993).
[b] USDA-NRCS (1998).

TABLE 1.3
Sizes of Soil Particle Classes

Class	Size
Sand	0.05–2 mm
Silt	0.002–0.05 mm
Clay	<0.002 mm (<2 μm)
Coarse fragments (not considered for soil texture analysis)	>2 mm

Source: Adapted from Soil Survey Division Staff. 1993. *Soil Survey Manual. USDA-SCS Agricultural Handbook 18*. U.S. Government Printing Office, Washington, DC.

feel sticky when rubbed. Sand and silt particles tend to be roughly spheroidal, with either smooth or rough edges. Clay particles are mostly flat and plate like; their large surface area influences soil chemical and physical characteristics. Notice that there is no such thing as a "loam" particle. "Loam" is the name for a mixture of particles of different sizes. The USDA defines 12 different combinations, called textural classes, for describing and classifying soils by texture (Figure 1.5). All percentages are on a dry weight basis.

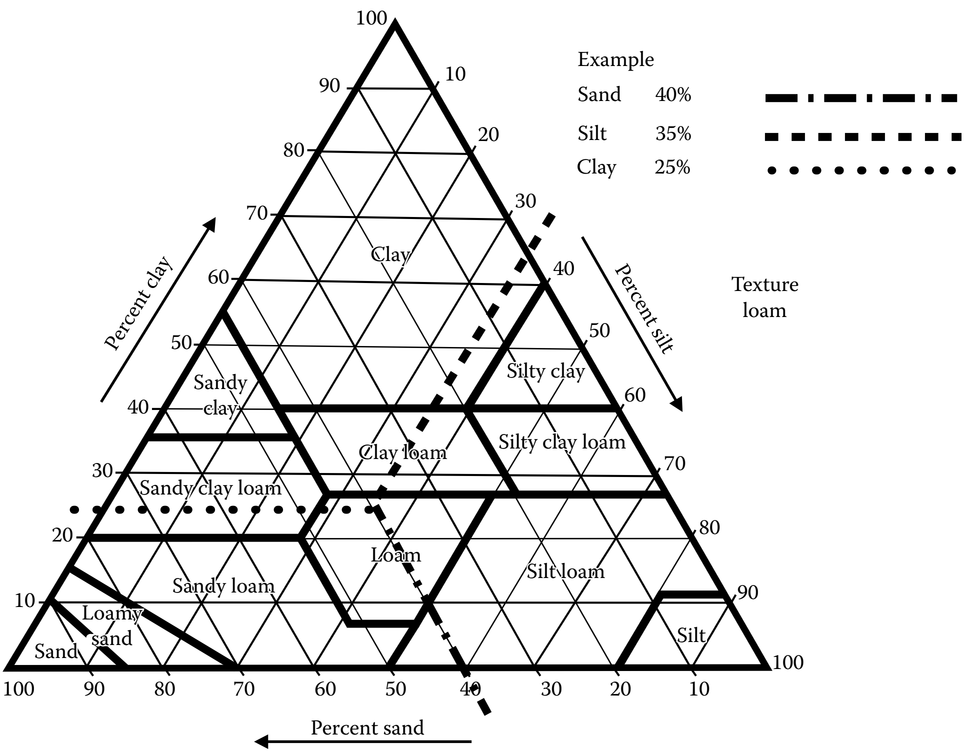


FIGURE 1.5
Soil texture triangle with an example of a loam soil sample. Read 40% sand-sized particles along the bottom axis from right to left and follow the 40% line upward at 60° to the left; "25% clay-sized particles" is read off the clay axis on the left side of the triangle, and "35% silt-sized particles" is read off the right axis. These three lines intersect in the "loam" area of the triangle; so, the sample has a loam textural classification.

Notice in Figure 1.5 that "sand," "silt," and "clay" are names of both individual particles and soil textures. If a soil sample is >90% sand- or silt-sized particles, the texture of the sample is named "sand" or "silt," respectively, after the dominant size fraction. However, less than half of the mass of a soil can be clay-sized particles and the material may still be called "clay"; this is because of the dominant influence of clay particles on the overall soil properties.

With training and practice, soil scientists can learn to estimate soil textures in the field by rubbing a moistened soil sample between their fingers and testing for properties such as ductility, grittiness, smoothness, stickiness, resistance to pressure, and cohesiveness. This art is locally specific because of regional variations in clay mineralogy. Wetland investigations for regulatory purposes, however, usually do not require the field accuracy necessary for soil mapping. Routine wetland investigations should record whether a soil horizon is generally sandy, silty, clayey, or loamy. This level of accuracy can be achieved by using a widely accepted flow chart for estimating soil textures (Figure 1.6), and can be documented by a disclaimer if the investigator lacks the training of a professional soil scientist.

The boundary between sandy and loamy hydric soil indicators (USDA-NRCS 2010; Chapter 8) is the boundary between loamy fine sand (sandy soil indicators) and loamy very fine sand (loamy and clayey soil indicators). A rule of thumb for determining whether the indicators for "sandy" or "loamy or clayey" soils should be used is to take a moist soil sample and roll it into a 1-in. ball. Drop the ball into the palm of your hand from a height of about 25 cm (10 in.). If no ball can be formed or if the ball falls apart when dropped, then, use the indicators for sandy soils. If the ball stays intact after dropping, use the indicators for "loamy or clayey" soils.

Mucky Mineral Textures

When the organic matter content of a mineral soil horizon is intermediate between organic and mineral soil materials, it is said to have a "mucky modified mineral texture," such as mucky sand or mucky sandy loam (Figure 1.2). These textures can only be learned by practicing with soil samples of known contents of clay and organic carbon.

Other Features

Formal soil descriptions include numerous distinctions in addition to those solicited on federal hydric soil field forms. Soil structure describes the aggregation of soil particles and the presence of large cracks and root channels. The terminology for soil structure is based on the concept of the natural soil aggregate (soil "ped") and its size, shape, and strength of expression. The details are available in standard soils texts and NRCS publications (Schoeneberger et al. 2012). Wetland delineation data sheets seldom require formal descriptions of soil structure, but often, redoximorphic features are found at horizon boundaries where water temporarily perches, such as at the contact where a horizon with a well-developed structure overlies a horizon with a minimal structure.

Other features that wetland delineators should be aware of when describing hydric soils include:

- Restrictive layers that cause abrupt changes in root density
- Compacted layers such as plow or traffic pans
- Different kinds of iron or manganese segregations
- External factors such as geomorphic position, water table depth, etc.

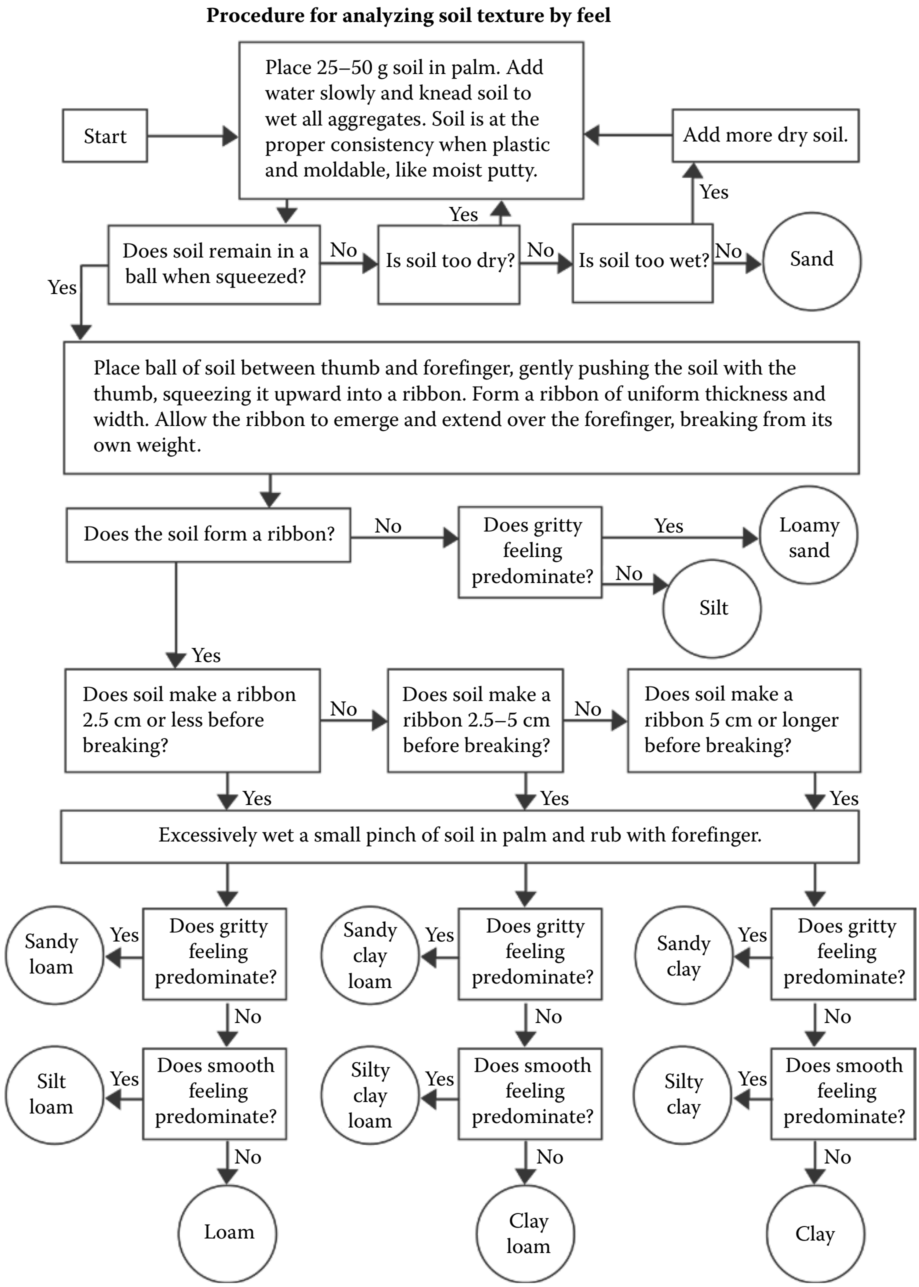


FIGURE 1.6
Flow chart for estimating soil texture by feel. To estimate soil texture, first wet the soil in the palm of your hand to its state of greatest malleability. It may take several minutes of manipulation to wet the smaller clay aggregates. If the soil gets too wet and puddles, just add more dry soil and rework it to optimum malleability. After the soil is adequately moistened, follow the flow chart by trying to make a ball and then a ribbon of the soil. A soil's ability to hold a ribbon shape reflects its clay content. Grittiness or smoothness of the ribboned soil indicates high content of sand or silt, respectively. (From Presley, D. R. and S. J. Thien. 2008. *Estimating Soil Texture by Feel.* Kansas State University Research and Extension, MF2852. http://www.ksre.ksu.edu/bookstore/pubs/MF2852.pdf)

The details of these features are described in professional soils publications (Vepraskas 2004; Schoeneberger et al. 2012).

Kinds of Soil Horizons

Soil scientists use the features discussed above to characterize individual soil horizons down through the soil profile; the major layers ("master horizons") recognized by U.S. pedologists (soil scientists) are O, A, E, B, C, and R horizons (Figure 1.1). Pedology distinguishes several varieties of each of the master horizons; the most significant of these subordinate horizons for the purposes of wetland science are listed in Table 1.4. Few soils have all the master horizons, and probably, no soil has all the subhorizons listed in Table 1.4. The wetland delineator usually inspects soil to 50 cm and, therefore, usually sees only the O and A horizons and the top of the E or B horizons, if present. A trained soil scientist, on the other hand, generally wants to investigate lower horizons as well, to understand the relation of the surface horizons to the landscape and its hydrology.

Organic horizons (O horizons) are typically associated with organic soils, or "Histosols," and an important subdiscipline of wetland and soil sciences involves the study and management of deep organic beds (see Chapter 10). Histosols generally have organic layers totaling at least 40 cm thick. Although organic horizons are not present in most mineral soils, when present as 1–2-cm-thick Oa horizons, they can be important hydric soil field indicators (Chapter 8). When they do occur in mineral soils, it is usually at the soil surface, unless the soil is buried by mineral matter washed in from flooding or upslope erosion.

In most soils, the uppermost mineral layer, or topsoil, is referred to as the A horizon. It is important to recognize the A horizon because hydric soils are usually identified from features immediately below it. The A horizon is usually the darkest layer in the soil (moist value/chroma darker than 4/2 in most hydric soil situations). It usually has more roots, organic matter, and biological activity than lower horizons and is more friable or crumbly. Most natural A horizons vary in thickness from approximately 5 to 30 cm, but some are thicker. Plowing may obscure A-horizon features because of mixing with subsoil materials. Plowed soil surfaces are referred to as "Ap" horizons. Ap horizons can be identified

TABLE 1.4

Subordinate Horizons of Greatest Significance to Wetland Science

Horizon	Significance
Oi	Fibric organic matter (little decomposition)
Oe	Hemic organic matter (intermediate decomposition)
Oa	Sapric organic matter (high decomposition)
Ap	Plowed A horizon
Bw	Weathering, weakly developed B horizon
Bt	Increase in illuvial clay in B horizon
Bg	Gleying significant
Btg	Increase in illuvial clay and significant gleying
Bh	Humus-rich subsoil, spodic horizon
W	Water layer within the soil. Wf is used for ice

by the abrupt, sharp lower boundary at the depth of a plow blade—generally 15–25 cm, depending on local agricultural practices.

The E horizon, when present, is a layer from which clay and iron oxides have been leached ("eluviated"). The E horizon is typically lighter in color than the rest of the soil above and below, usually gray to white. This light color is due to the fact that the loss of clay, iron, etc. leaves a concentration of sand and silt particles of quartz or other resistant minerals. It is important to recognize E horizons because their low chroma and high value can be mistaken for evidence of wetness and Fe reduction. While many wet soils can and do contain E horizons, it is important to determine if they are truly a product of wetness, or if they are a product of loss of clay and/or Fe. E horizons of hydric soils typically contain redox concentrations (i.e., reddish mottles). E horizons are underlain by a layer having a higher content of clay (Bt horizon) or transported organic material (Bh horizon).

The B horizon is the layer of most obvious mineral weathering and is the layer into which the material translocates from the overlying E and A horizons and accumulates. The B horizon has soil peds (coherent aggregates) unless the soil is nearly pure sand. Upland B horizons have the colors (generally browns) of the iron minerals that weather out of the original parent material. Wetland B horizons are grayer due to reduction and removal of iron-pigmenting minerals (see Bg horizons below and Chapter 7).

Most hydric soils have a subsoil horizon that is seasonally anaerobic due to high water tables and chemical reduction. This is termed a Bg horizon; the "g" indicates processes of "gleying," that is, chemical reduction of iron or manganese. Matrix colors of Bg horizons are usually gray, with chromas of 2 or less and values of 4 or more, usually with redox concentrations (reddish mottles); Bg colors are not restricted to the Munsell gley charts. Not all Bg horizons are indicative of hydric soils; for example, deeper water tables may create Bg horizons below a depth of 30 cm, which is not shallow enough for the soil to be hydric.

Bt horizons are zones where clay accumulates from above ("illuviates"), often from an E horizon. The increased clay content is significant to hydric soils because water can perch in and on top of Bt horizons and cause redoximorphic features to form. Such perched water, however, may not be present long enough or frequently enough to cause the formation of gray matrices or hydric soils.

Bh horizons are the dark subsoil horizons often found in sandy and loamy soils under coniferous vegetation, especially in the Southeast Coastal Plain and in glacial outwash plains (Spodosols). Their morphology is distinctive: they almost always underlie a white E horizon; they are black to dark reddish brown; and their boundaries with horizons above and below are usually very sharp. The *Regional Supplements to the Corps of Engineers Wetlands Delineation Manual* (U.S. Army Corps of Engineers 2010: http://www.usace.army.mil/missions/civilworks/regulatoryprogramandpermits/reg_supp.aspx) recognizes Spodosols as problem soils with blotchy color patterns that can be mistaken for iron concentrations induced by anaerobic conditions.

The geologic material in which soils form is termed the C horizon, if unconsolidated, or R horizon, if it is bedrock. Many soils in fluvial settings have only an A and a C horizon, entirely lacking O, E, and B horizons. C horizons retain the structure and color of the original parent material. In a fluvial setting, the C horizon would retain evidence of sedimentary stratification, whereas B horizons in the same setting would have developed enough structure that the boundaries between depositional strata are obliterated. Few wetlands have an R horizon because most depressional areas are deeper than 2 m to bedrock.

A W layer is a relatively new addition to horizon nomenclature, and is used to denote the presence of a layer of water within or beneath the soil. The water layer is designated as Wf if it is permanently frozen, and as W if it is not permanently frozen. The W or Wf designation is not used for shallow water, ice, or snow above the soil surface (Lindbo et al. 2008).

Soil Taxonomy

Soil Taxonomy (Soil Survey Staff 1999) is the most comprehensive classification system used to catalog soils in the United States. Wetland scientists need to be familiar with the highest level of the system and with a handful of lower-level taxa subordinate distinctions to understand concepts and terminology in the hydric soils literature.

Soil Taxonomy is a hierarchical classification system with six levels (order, suborder, great group, subgroup, family, and series; see Table 1.5). The highest level is composed of 12 soil orders (Table 1.6); soil orders are based on fundamental differences in soil genesis. The second level, the suborder, often indicates the soil moisture regime of the soil or its annual precipitation inputs. Sometimes, the third level (great group) and often the fourth level (subgroup) carry information about soil hydrology. All four levels are communicated in the taxonomic name. The fifth level (family) provides information about soil texture and mineralogy, among other things. The sixth level is the soil series name, for example, "Sharkey" or "Myakka" or "Wakeley." These series names can be thought of as comparable to the binomial species name in the Linnaean classification systems of plants and animals. As of 2013, approximately 21,800 soil series (i.e., different usefully mappable types of soil) were recognized in the United States. Most soil maps in the United States include distinctions between soil types and phases, which are subsets of the series, much as varieties are subsets of plant or species.

TABLE 1.5

Hierarchy of Soil Taxonomy and Example Using Tonka Soil Series

Level	Distinctions	Example	Significance
Order	Major soil-forming processes	Mollisol	Thick, dark surface + base saturation required
Suborder	Moisture regime, parent material, and secondary processes	Alboll	Albic (E) horizon
Great group	Diagnostic layers, base status, horizon expression, and water perching	Argialboll	Clay accumulation in Argillic horizon
Subgroup	Moisture regime refinements	Argiaquic Argialboll	Poorly drained
Family	Texture, mineralogy, cation exchange capacity, temperature, and acidity	Fine, smectitic, and frigid Argiaquic Argialboll	High clay content, expansive clay, and cold temperatures
Series	Comparable to species in plant taxonomy	Tonka	
Phase	Slope, flooding, surface texture, etc.	Tonka silt loam	Surface horizon is silt loam

Note: The complete family name of the Tonka series is "Fine, smectitic, frigid Argiaquic Argialbolls."

TABLE 1.6

Soil Orders (Highest Level of *Soil Taxonomy*)

Order	Suffix in Taxonomic Name	Significance	Typical Location
Alfisols	-alf	Significant clay illuviation and high base status	Cool, humid deciduous forests
Andisols	-and	Significant presence of volcanic glass	Areas of volcanic ash deposition
Aridisols	-id	Desert climate	Deserts
Entisols	-ent	Minimal soil development	Sands; recent deposits
Gelisols	-el	Permafrost	High latitude and elevation
Histosols	-ist	Formed in deposits of organic material	Wet closed depressions
Inceptisols	-ept	Young soil with incipient development	Active landforms nationwide
Mollisols	-oll	Thick, dark A horizons	Prairies
Oxisols	-ox	High content of iron oxides	Tropics
Spodosols	-od	Subsoil horizon of humus and Al/Fe sesquioxides	Humid coniferous forests
Ultisols	-ult	Significant clay illuviation and low base status	Warm, humid deciduous forests
Vertisols	-ert	Shrink/swell activity due to clays	Clay beds, especially south-central United States

TABLE 1.7

Words and Phrases from Soil Taxonomy That Have Particular Significance to Wetland Science

Word or Phrase	Meaning
Aqu-	An aquic (or seasonally reducing) moisture regime (e.g., Aqualf). Soils with a different syllable in the suborder (second) position have drier moisture regimes (e.g., Udalf).
Epi- versus endo-	A perched water table (e.g., Epiaqualf) in contrast with a water table rising from the bottom of the soil (Endoaqualf).
Aeric	Somewhat ameliorated wetness limitations (e.g., Aeric Epiaqualf). The water table is within 75 cm of the soil surface.
Histic	High organic matter content in the soil surface and is usually formed under extreme wetness (e.g., Histic Endoaquoll).
Mollic taxa and Mollisol order (suffix is "oll")	Thick, dark A horizons, which make hydric soil identification difficult because redoximorphic features tend to be masked by organic matter to a considerable depth (e.g., Mollic Natrustalf; and Typic Endoaquoll).
Fluv-	Alluvial deposition; possible flooding hazard (e.g., Fluvaquent).
Vertic taxa and Vertisol order (suffix is "ert")	High clay contents with high shrink–swell capacity; hydrologic inputs are usually surficial rather than from below (e.g., Vertic Epiaquept; and Aeric Epiaquert).

The connotative translation code for the constituent parts of soil names is found in *Soil Taxonomy* (Soil Survey Staff 1999) and in many soil textbooks. Periodically, Keys to Soil Taxonomy is published to update users on recent changes in the classification system (12th edition, 2014). Most of the distinctions in *Soil Taxonomy* are not significant to hydric soils work; the most pertinent are listed in Table 1.7.

Soil Series within Landscapes

Soils and water tables often lie in landscapes in continua descending from dry soils with deep water tables down to wet soils with shallow water tables. Such wetness gradients are variously referred to as toposequences or drainage catenas, *catena* being the Latin word for a chain consisting of separate links. Wetlands and hydric soils typically occur at the lowest position in the landscape, receiving surficial runoff and subsurface drainage. Chapter 3 discusses the variety of other landscapes where wetlands and hydric soils develop at other landscape positions, belying the layman's topographic stereotypes.

Figure 1.7 shows a block diagram of a glacial landform in Barnes County, ND, with a specific soil series occupying specific positions down the drainage catena. Tonka typically has hydric soil morphology; Barnes, Buse, and Svea typically are better drained. These soils may be considered to form a drainage catena if they frequently are found associated with each other in such topographic relationships.

For decades, Soil Survey has referred to soils with similar hydrologic regimes using the shorthand of natural drainage classes. In this example, the drainage catena is Barnes (well drained), Svea (moderately well drained), Hamerly (somewhat poorly drained), and Tonka (poorly drained soils). The parenthetical, comparative shorthand terms can be used appropriately within specific drainage catenas but have been carelessly abused over time to suggest that other soil series in other landscapes—and even other regions and other climates—have comparable hydrologic regimes just because they are labeled with the same drainage class name, despite fundamental differences in geomorphology and the hydrologic regime. To compound matters, within states or geomorphic regions, contradictions have developed when hydric soils *per se* have been exclusively equated with poorly and very poorly drained classes. The drainage class terminology will continue to be used unofficially because it is so useful to make comparisons within specific landscapes, but the Soil Survey Program is beginning to abandon formal use of the classification system.

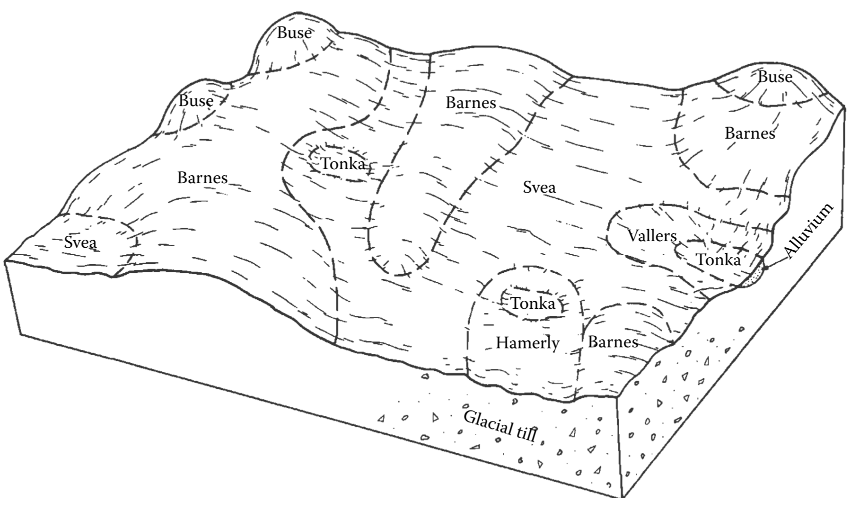


FIGURE 1.7
Schematic of landscape positions for different natural drainage classes in southeastern North Dakota. (From Opdahl et al. 1990. *Soil Survey of Barnes County, North Dakota.* USDA-SCS in cooperation with the North Dakota Cooperative Extension Service and North Dakota State Soil Conservation Committee, Agricultural Experiment Station, U.S. Government Printing Office, Washington, DC.)

Soil Survey

The National Cooperative Soil Survey is the United States' program to map the soils of the nation, their distribution, properties, and potentials and limitations for land use. The fundamental concept of the United States soil survey is the soil map unit. The map unit is an abstract concept describing the kinds of soils generally mapped together. In this regard, soil mapping is analogous to vegetation mapping. The legend of a vegetation map may include a map unit of "Red Oak Forest." Not all plants within areas so mapped are R oaks; similarly, for example, not all soils within areas mapped as "Sharkey clay" are Sharkey soils. In most cases, the dominant plant species or soil series within the map unit is the one after which the unit is named, but there can be numerous inclusions of other plants or soils.

Most soil maps in the nation in areas with a history of agriculture were made by second-order surveys (scales of 1:12,000–1:30,000). The minimum size delineation on a second-order map is 0.6–4 hectares (depending on scale), and most map delineations are considerably larger because of constraints on map legibility. First-order soil maps cover a smaller land area, are more detailed, and usually are produced for a particular project. Third- and fourth-order maps cover larger land areas and are less detailed. It is not recommended to make site-specific hydric soil determinations from the office using second-, third-, or fourth-order soil survey information alone because of the presence of inclusions within soil-mapping units. On-site investigations are required. Also note that most soil maps were not made with hydric soils in mind. The concept of a hydric soil was developed after the majority of the nation's land had already been mapped.

Second-order soil surveys map soils at the level of the soil phase, which is a subset of the soil series. Typical distinctions made at the level of the soil phase are slope gradient, flooding frequency, and surface texture. Many soil series have both hydric and nonhydric phases, even within the same county. Take, for example, two neighboring soils in Levy County, Florida, both of them dominated by the Myakka soil series (USDA-NRCS 2013). Map unit 37 is the phase "Myakka mucky sand, occasionally flooded" and is predominately hydric (95% hydric soil); map unit 38 is the phase "Myakka sand" and is predominately nonhydric (6% hydric soils).

As of this chapter, the USDA Soil Survey Program has transitioned to Internet-based soil map dissemination (Web Soil Survey, websoilsurvey.nrcs.usda.gov) and is no longer distributing the original paper soil survey reports, published pursuant to completion of individual survey projects. Soil Survey data are now updated annually; so, the old paper reports are obsolete, though the narrative portions of the text are being archived on the Web Soil Survey website. The Soil Survey Program currently is correcting artifacts left over from a half century of mapping at the county/parish level, including in wetlands.

Summary

Wetland soil investigations utilize the same or abbreviated protocols used for standard soil survey projects. If the study is limited to hydric soils determinations, it usually suffices to describe horizon depths, color, redoximorphic features, and textural class. In mineral soils, many hydric soil determinations are made below the A horizon, usually in the B horizon; however, be aware of alternative horizons and features that may be present at these shallow depths. Some hydric soils must be determined from features composed of soil organic matter. Prior to an on-site investigation for any purpose, it is useful to consult Web Soil Survey. The appropriate use of soil survey products, however, requires familiarity with soil-mapping conventions, including map scale, the concept of drainage catenas, map unit inclusions, and terms in *Soil Taxonomy* that apply to soil wetness.

Acknowledgments

The permission to publish this chapter has been granted by the USDA Natural Resources Conservation Services.

References

Kriz, D. M. 2007. List of hydric soils by county. In V. C. Carlisle (Ed.) *Hydric Soils of Florida Handbook*, 4th ed. Florida Association of Environmental Soil Scientists, Gainesville, FL, pp. 155–409.

Lindbo, D., R. Miles, D. Presley, and N. Ransom. 2008. Soil profile description. In S. Logsdon, D. Clay, D. Moore, and T. Tsegaye (Eds.) *Soil Science, Step-by-Step Field Analysis*. Soil Science Society of America. Madison, WI.

Lipscomb, G. H. 1992. *Soil Survey of Monroe County, Pennsylvania.* USDA-SCS in cooperation with the Pennsylvania State University and Pennsylvania Department of Environmental Resources, U.S. Government Printing Office, Washington, DC.

Lueth, R. A. 1974. *Soil Survey of Hennepin County, Minnesota.* UDA-SCS in cooperation with the Minnesota Agricultural Experiment Station, U.S. Government Printing Office, Washington, DC.

Munsell Color. 2013. *Munsell Soil Color Charts*. X-Rite Corporation, Grand Rapids, Michigan.

Presley, D. R. and S. J. Thien. 2008. *Estimating Soil Texture by Feel*. Kansas State University Research and Extension, MF2852. http://www.ksre.ksu.edu/bookstore/pubs/MF2852.pdf

Schoeneberger, P. J., D. A. Wysocki, E. C. Benham, and Soil Survey Staff. 2012. *Field Book for Describing and Sampling Soils, Version 3.0*. Natural Resources Conservation Service, National Soil Survey Center, Lincoln, NE.

Soil Science Society of America. 2008. *Glossary of Soil Science Terms.* Soil Science Society of America, Madison, WI.

Soil Survey Division Staff. 1993. *Soil Survey Manual. USDA-SCS Agricultural Handbook 18.* U.S. Government Printing Office, Washington, DC.

Soil Survey Staff. 1999. *Soil Taxonomy: A Basic System of Soil Classification for Making and Interpreting Soil Surveys*, 2nd ed. *USDA-SCS Agricultural Handbook 436*. U.S. Government Printing Office, Washington, DC.

Soil Survey Staff. 2014. *Keys to Soil Taxonomy*, 12th ed. USDA-NRCS, Washington, DC.

United States Department of Agriculture, Natural Resources Conservation Service. 2010. In L. M. Vasilas, G. W. Hurt, and C.V. Noble (Eds.) *Field Indicators of Hydric Soils in the United States, Version 7.0*. USDA, NRCS, in cooperation with the National Technical Committee for Hydric Soils. Available at: ftp://ftp-fc.sc.egov.usda.gov/NSSC/Hydric_Soils/FieldIndicators_v7.pdf

U.S. Army Corps of Engineers. 2010. *Regional Supplements to Corps Delineation Manual*. http://www.usace.army.mil/missions/civilworks/regulatoryprogramandpermits/reg_supp.aspx, accessed December 7, 2013.

USDA-NRCS. 2013. Web soil survey. Available at: http://websoilsurvey.sc.egov.usda.gov/App/HomePage.htm

Vepraskas, M. J. 2004. *Redoximorphic Features for Identifying Aquic Conditions*. North Carolina Agricultural Research Service Technical Bulletin 301. North Carolina State University, Raleigh, NC.

2

History of the Concept of Hydric Soil

Michael J. Vepraskas

CONTENTS

Introduction

The concept and definition of hydric soils, the soils found in wetlands, originated in 1979 and evolved over approximately a 20-year period as the needs for hydric soil changed from verifying wetland maps to placing boundaries around specific wetlands in the field. The term "hydric soil" was used for the first time in the wetlands classification system of Cowardin et al. (1979, p. 3) for the soils found in natural wetlands. The Cowardin classification system was developed to assist in the *National Wetlands Inventory* (NWI) of the U.S. Fish and Wildlife Service that was a national wetlands-mapping program. It defined wetlands as "…*lands transitional between terrestrial and aquatic systems where the water table is usually at or near the surface or the land is covered by shallow water. For purposes of this classification wetlands must have one or more of the following three attributes: (1) at least periodically, the land supports predominantly hydrophytes, (2) the substrate is predominantly hydric soil, and (3) the substrate is nonsoil and is saturated with water or covered by shallow water at some time during the growing season of each year.*"

The NWI survey was to be done using aerial photos rather than "on the ground" surveys. The U.S. Department of Agriculture's (USDA) soil survey maps were going to be used to check the accuracy of the NWI's wetland delineations, because it was felt that the soil maps would show the locations of the "hydric" soils (Mausbach and Parker 2001). The NWI asked the USDA Soil Conservation Service (SCS) (forerunner to the Natural Resources Conservation Service, NRCS) to supply a list of names of the hydric soils that were shown on the soil maps. This was easily said than done, because the USDA did not map hydric soils per se, and had no procedures in place for identifying them.

Preparing a list of hydric soil names for the NWI program was the major focus of the USDA soil scientists who worked with hydric soils from approximately 1979 to 1985. After that time, on-site identification of hydric soils became increasingly more important as the boundaries of jurisdictional wetlands needed to be drawn on maps to enforce wetland protection laws. The objective of this chapter is to trace the changes in the concept of *hydric soil* by focusing on how the definition of this soil group has evolved to meet these changing needs.

Developing a List of Hydric Soils

To support the NWI mapping efforts, the USDA's SCS was given the tasks of defining hydric soils, establishing criteria for their identification, and preparing a list of hydric soils that could be used along with soil maps to evaluate the accuracy of the NWI maps (Mausbach and Parker 2001). A team of USDA soil scientists and biologists decided to begin by visiting natural wetlands with established hydrophytic vegetation. A definition of hydric soil was needed to identify areas to be visited, because the properties of hydric soils were not yet known. The first definition devised is shown in Table 2.1 for 1977. It was very simple and emphasized hydrology and vegetation, but no soil properties were mentioned. The definition implies that a hydric soil could be identified by either hydrology or

TABLE 2.1

Definitions of Hydric Soils That Have Been Used by USDA Soil Scientists to Prepare Lists of Hydric Soils

Date	Definition	Comments
1977	Hydric soils are soils with water at or near the surface for most of the growing season or the soil is saturated long enough to support plants that grow well in a wet environment.	First definition, used by selected field personnel for testing
1980	Hydric soils are soils that have reducing conditions for a significant period of the growing season (soil is virtually free of oxygen) in the major part of the root zone and are saturated (with free water) within 25 cm of the surface. Most hydric soils have properties that reflect dominant color in the matrix as follows: (1) if there is mottling, the chroma is 2 or less, and (2) if there is no mottling, the chroma is 1 or less.	Second definition, sent to USDA personnel in states to begin to assemble lists of hydric soils
1985	A hydric soil is a soil that is saturated, flooded, or ponded in its undrained condition long enough during the growing season to develop anaerobic conditions that favor the growth and regeneration of hydrophytic vegetation.	First definition published by the NTCHS
1986	The NTCHS modified the definition by deleting "in its undrained condition."	Definition used in the Corps of Engineers *Wetlands Delineation Manual of 1987*
1987	A hydric soil is a soil that is saturated, flooded, or ponded long enough during the growing season to develop anaerobic conditions in the upper part.	Definition published with the second edition of *Hydric Soils of the United States*
1994	A hydric soil is a soil that formed under conditions of saturation, flooding, or ponding long enough during the growing season to develop anaerobic conditions in the upper part.	Current definition, published in Federal Register

vegetation. Because it was developed for "unaltered" wetlands that were not drained, the definition required that soils be saturated "at or near the surface" for some period of the year. The hydrology component in the definition was emphasized to exclude soils that had been drained for agriculture, protected by levees along rivers, or that had been made drier by climatic changes or river down cutting. When hydrology could not be observed at a site, then, the mere presence of hydrophytic vegetation was apparently sufficient to identify the hydric soil. From 1977 to 1980, the USDA scientists visited wetlands around the United States that met the definition of 1977 to describe and classify the soils they found.

Using the information gleaned from 3 years of field investigations, the USDA scientists developed the 1980 definition of hydric soils shown in Table 2.1. The intent was to have a definition that could be sent to USDA state offices where local soil scientists would prepare the lists of hydric soils for their state. The hydrologic requirements in the 1980 definition stressed that hydric soils had to be saturated during the growing season and this excluded drained soils from being hydric. Reducing conditions (anaerobic conditions) were required because these were necessary for hydrophytic plants to predominate on the soil, as well as to form the gray soil colors found in hydric soils (Vepraskas 1994; Chapter 7). Field investigations showed that the soil color requirements (based on Munsell chroma) that were characteristic of hydric soils were visible within 25 cm of the soil surface.

The changes to the 1977 definition were striking because now, the focus was almost entirely on soil characteristics that were used to classify soils in *Soil Taxonomy,* which is the name of the soil classification system used by the USDA (Soil Survey Staff 1975).

Soils with an aquic moisture regime become reduced (anaerobic) after being saturated by either groundwater or the water of the capillary fringe (Soil Survey Staff 1975). This concept is reflected in the definition of 1980 by hydric soils now having "….reducing conditions (soil is virtually free of oxygen) within 25 cm of the surface)." It was also assumed that most roots of hydrophytic plants would occur within 25 cm of the soil surface (Mausbach and Parker 2001). At this point in time, all hydric soils were assumed to be saturated and reduced during most years, drought years being the exceptions. No minimum period of saturation was specified. In practice, aquic moisture regimes were identified on the basis of soil color, because measurements of saturation or reducing conditions were generally not made in most soils. The soil characteristics that were used to identify aquic moisture regimes were generally the low (≤2) chroma colors (gray colors) and mottling that usually consisted of accumulations of iron oxides.

The hydric soil definition of 1980 was forwarded to USDA state offices so that the local soil scientists could assemble state lists of the soils that met the definition. Soils on the lists were to be identified by their soil series name. The lists of soil series from the states would then be compiled into one national list that could be used, along with soil maps, to evaluate the NWI maps. The soil series is a name given to a group of soils that have similar soils horizons, soil chemical and physical properties, and classification. Soils within the same series can differ in the texture of the surface horizon.

The results received from the states showed that this approach would not work easily. Soil series could not be consistently classified as "hydric" or "not hydric" using the hydric soil definition of 1980. A soil that was considered to be a hydric soil in one state might not be considered hydric in another (Mausbach and Parker 2001). The reason for this is that in some cases, all the soils within the same series do not meet the requirements for a hydric soil. Soils within a given series can be subdivided into soil *phases* where some will meet the proposed hydric requirements and some will not. For example, a series might have both drained and undrained phases, where only the undrained soils would meet the hydric requirements as defined. Flooded and not-flooded phases are also found within some series where only the flooded phase would meet the hydric soil requirements. Unless the soil phases were identified, there would not be a consistent interpretation among soil scientists in different states as to whether a given series contains hydric soils or not.

Another problem with using a simple definition to identify names of hydric soils was that the depth ranges needed for saturation were much shallower than those currently used in *Soil Taxonomy.* Although periods of saturation and reducing conditions were required within 25 cm of the surface for a soil to be hydric, and presumably to have an aquic moisture regime, the aquic moisture regime as defined in *Soil Taxonomy* required saturation and reduction within the upper 40–50 cm of the soil (Soil Survey Staff 1975). This meant that many soils with an aquic moisture regime would span the wetland–upland boundary and in some cases, might not be in a wetland at all. In addition, measurements of saturation and reducing conditions are usually not made on soils because they are simply too expensive. Instead, soil colors are used to infer that saturated and reduced conditions have occurred, but the standards for doing this vary from state to state. For example, a soil that had low chroma (gray) colors in more than 20% of a horizon within 25 cm of the surface could be considered to be saturated long enough to qualify as a hydric soil in one state, while in another state, this soil might be considered too "dry." In summary, this approach for developing a list of hydric soils was found to be unworkable, and another strategy needed to be developed.

The USDA soil scientists concluded that a national list of hydric soils had to be developed by using a set of specific *criteria* that included soil classification information and also other requirements for depth of saturation as well as requirements for flooding and ponding. This comprehensive set of criteria would then be used to search the USDA's national database of soils that were mapped in the United States. The new criteria for hydric soils needed to be defined by a group of scientists who had extensive experience with wetland identification.

The National Technical Committee for Hydric Soils Takes Over

In 1981, the USDA established the *National Technical Committee for Hydric Soils* (NTCHS) to oversee the definition of hydric soils as well as to develop a list of these soils. This group included soil scientists and a biologist from the USDA, a biologist with the U.S. Fish and Wildlife Service, and two wetland soil scientists from universities. The original duties of the NTCHS are listed in Table 2.2. At this point in time, developing a list of hydric soils

TABLE 2.2

Duties Assigned to the NTCHS

Develop the definition and criteria for hydric soils
Develop procedures for reassessing the criteria and the list of hydric soils
Develop an operational list of hydric soils and distribute it to SCS state offices and cooperators
Coordinate activities with the National Wetland Plant List Review Panel
Provide continuing technical leadership in the formulation, evaluation, and application of criteria for hydric soils

Source: After Mausbach, M. J. and W. B. Parker. 2001. Background and history of the concept of hydric soils. In J. L. Richardson, and M. J. Vepraskas (Eds.) *Wetland Soils: Genesis, Hydrology, Landscapes and Classification*. Lewis Publishers, Boca Raton, FL.

remained the major focus for the USDA and the NTCHS. However, in 1983, representatives of the U.S. Army Corps of Engineers and the U.S. Environmental Protection Agency were added to the NTCHS because of the growing need to use hydric soils for enforcement of the Clean Water Act (National Research Council 1995).

At the time the NTCHS was formed, the concept of a hydric soil still required these soils to support the growth of hydrophytic vegetation. This meant that during most years, hydric soils had to be saturated and reduced (anaerobic) for some period of time (Mausbach and Parker 2001). Soils drained for agriculture or protected from flooding by levees or dams were not considered to be hydric soils. This concept, while reasonable in principle, made assembling a national list of hydric soils difficult, because a soil's classification could not be used to determine if a soil was in fact a hydric soil. The USDA's soil scientists who were completing a soil survey were classifying soils based on their "natural state." Draining a soil did not affect its USDA soil classification because in its natural state, a drained soil would still be considered to have an aquic moisture regime even though the soil may no longer experience saturation within 25 cm of the surface. The rationale for drainage not affecting a soil's classification was twofold. First, if the drainage system was not maintained (e.g., ditches became plugged or were filled), then, the soils would again saturate to within 25 cm of the surface. The same would hold true for soils protected by levees that were not maintained or failed as in the case of the levees protecting New Orleans, LA following Hurricane Katrina in 2005. Drained soils also retained their original gray colors and so, the morphology of the soil could not be used to separate drained from undrained soils. The second reason why drainage was not reflected in the classification was because USDA soil scientists who mapped and classified soils had no way of knowing the impact of the drainage system at a given mapping location (Smith 1986). There was neither time nor money to allow monitoring wells to be installed to determine the depth to which a drained soil would saturate.

Because of the difficulty in identifying hydric soils on soil maps if the soils had to be "undrained," the NTCHS moved in 1985 to define hydric soils as needing to be saturated and anaerobic only in their "natural state" (Table 2.1). The definition of hydric soils was changed to state that in its *undrained condition* "the soil is saturated, flooded, or ponded long enough during the growing season to develop anaerobic conditions…." In essence, this meant that if a soil had formed in a wetland, and had developed the soil color characteristics that were typical of hydric soils in wetlands, then, it would be considered a hydric soil even if the soil had been drained. With this change in concept, the NTCHS was able to develop a procedure for preparing a list of hydric soils using USDA records of the soils that had been mapped.

The USDA has a database that includes information on all soil series that have been mapped in the United States by the National Cooperative Soil Survey. The database, called the Soil Interpretations Record (SIR), contains the following information for every soil series mapped: soil classification, soil physical and chemical properties, drainage class, estimated water table depths, and the frequency and duration of either flooding or ponding. To identify the soils in the database that were likely to be hydric, a computer program was developed to search the database using only the soil data that were in the database. While this seems obvious, the point is that the hydric soil criteria were not developed to be the optimum way to either identify hydric soils or to define them. The criteria had to be developed around the available information in the SIR.

Several different versions of hydric soil criteria were developed and tested (Mausbach and Parker 2001). In 1985, the first workable set of criteria was defined for identifying hydric soils in the SIR database, and these are shown in Table 2.3. Criterion 1 refers to organic soils

TABLE 2.3

Hydric Soil Criteria Used as Part of the Definition of Hydric Soils in 1985

1. All Histosols except Folists, or
2. Soils in Aquic Suborders, Aquic Subgroups, Albolls Suborder, Salorthids Great Group, or Pell Great Groups of Vertisols that are
   a. Somewhat poorly drained and have a water table less than 0.5 ft from the surface at some time[a] during the growing season, or
   b. Poorly drained or very poorly drained and have either
      i. A water table at less than 1.0 ft from the surface at some time[a] during the growing season if the permeability >6.0 in./h in all layers within 20 in., or
      ii. A water table at less than 1.5 ft from the surface at some time[a] during the growing season if the permeability is less than 6.0 in./h in any layer within 20 in., or
3. [b]Soils that are ponded during any part of the growing season, or
4. Soils that are frequently flooded for a long duration or a very long duration during the growing season.

[a] In 1991, "at some time" was changed to "a significant period (usually more than 2 weeks)."

[b] In 1991, criterion 3 was changed to soils that are frequently ponded for a long duration or a very long duration during the growing season.

(i.e., peats and mucks with organic layers of 41 cm or more thick) that are called Histosols in *Soil Taxonomy*. Most Histosols were inundated or saturated to the surface during their formation, and the reduced conditions that developed following saturation slowed decomposition of the organic tissues. Folists are a special, and rare, kind of organic soil that forms under aerobic conditions due to excessive amounts of leaves falling on rock or very gravelly soil material (Soil Survey Staff 1975). They were excluded from being considered as hydric soils because they do not form in saturated environments.

Criterion 2 consisted of several parts and identified the mineral soils believed to have water tables that remained close to the surface during the growing season. All the soil classifications shown had aquic moisture regimes. Three natural soil drainage classes were used in the definitions, as were water table depths during the growing season, and in some cases soil permeability. It was assumed that hydric soils would have to be saturated within 25 cm of the surface during the growing season to support the growth of hydrophytic plants. Somewhat poorly drained soils are generally not saturated near the surface long enough to occur in wetlands. Exceptions do occur and it was believed that somewhat poorly drained soils could occur in wetlands if the water table was very shallow or less than 15 cm for short periods of time. Poorly and very poorly drained soils were included as one group with two subcategories that differed by water table depth and permeability. The soils found in wetlands usually occur in these two drainage classes. A shallower water table depth (<30 cm) was required where permeabilities in soil layers were >6 in./h (>15 cm/h, e.g., sandy soils) because these soils would drain quickly after rains and saturation of the surface horizons would probably only occur where the water tables were close to the surface. These soils also have relatively thin capillary fringes and it was believed that this might also maintain anaerobic or reduced conditions above the water table (Mausbach and Parker 2001). A deeper water table (<1.5 ft, or 45 cm) was included where permeabilities in soil layers were <6 in./h (<15 cm/h, e.g., loams and clays) because such soils would drain slowly after rains keeping the surface saturated for extended periods. It was generally believed but not proven that these soils would also have thick capillary fringes that could keep the soil layers anaerobic above the water table.

Criteria 3 and 4 dealt with inundated soils whose surfaces were covered either by moving water (flooded soils) or stagnant water (ponded soils). The soils meeting these criteria do not necessarily have aquic moisture regimes with seasonally high water tables, and may not occur in the three drainage classes included in criterion 2. Such soils could be found in wetlands if the flooding or ponding occurred frequently for a long duration during the growing season (Mausbach and Parker 2001).

Evolution of the Hydric Soil Criteria and the List of Hydric Soils

Using the criteria for hydric soils (Table 2.3), a list of soil series names that had phases which were hydric soils was compiled and published as the national list of *Hydric Soils of the United States 1985* (USDA 1985). This list of soil names was to be used along with soil maps to determine if an area of land could contain hydric soils. To use the list, a person would find a piece of property on a soil map and then identify the name of the *soil map unit* that the property was in. The soil map unit was the area whose boundaries were drawn or delineated on a soil map. A cross section of a Pungo map unit is shown in Figure 2.1. The map unit name (e.g., Pungo muck, 0%–1% slopes) contained the name of the series for the dominant soil (usually occupying 85% or more of the map unit) as well as other properties such as texture of the topsoil and land slope. If the map unit name was on the national list of hydric soils, then, the soil map unit would very likely contain hydric soils. A second edition to the list of hydric soils was published in 1987 that was very similar to the first edition but with a revised definition of hydric soils (USDA 1987).

Having a national list of hydric soils that was based on the names of soil series and phases identified through a data base search proved to contain inaccuracies over time, and was difficult for users who were not soil mappers to use reliably along with soil maps.

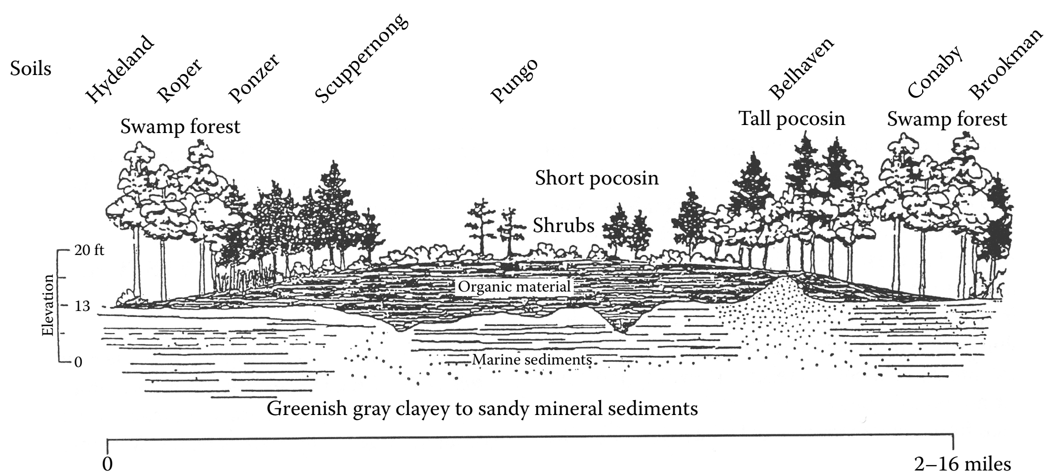


FIGURE 2.1
Variation in soil series across a wetland in the North Carolina Coastal Plain. If the area was drawn on a soil map and described with the soil map unit named as Pungo for the major soil map unit component, it is likely that one or more of the other components would be present due to variation in thickness of the organic material. (From Daniels, R. B. et al. 1999. *Soil Systems in North Carolina*. Soil Science Department, North Carolina State University, Raleigh, used with permission.)

Difficulties occurred because in many cases, a map unit (area delineated on the soil map) contained more than one kind of soil, with each different soil being a *map unit component* within the entire map unit. The name of the map unit component occupying the largest area was used to name the map unit. The other map unit components, which were usually members of a soil series different from the dominant component, commonly occupied <15% of the map unit. These minor components may or may not be hydric soils. In some cases, map units would have a dominant component that did not meet one of the hydric soil criteria, and whose series name was not on the list of hydric soils, while the minor components did meet one of the hydric soil criteria and whose series name was on the list of hydric soils.

After 1991, the hydric soil list was modified to make it more accurate, easier to compile, and easier to use with soil maps (USDA, 1991). Instead of simply listing the names of soil series that met one of the hydric soil criteria, the list was changed to show the names of soil map units that contained components that were hydric soils. The names of the hydric components were also included on the list. In this way, a user went directly from a soil map to the list of hydric soils.

Federal and State Wetland Protection Laws Influence Hydric Soil Concepts

In the 1980s, laws were enacted that protected wetlands from being filled in or drained. Two wide-ranging federal acts that were responsible for protecting wetlands were the Clean Water Act and the Food Security Act (National Research Council 1995). The Clean Water Act essentially focused on wetlands that were not on agricultural land, while the Food Security Act focused on wetlands on agricultural lands. The history of these acts was reviewed by the National Research Council (1995). These laws changed the direction in how hydric soils were used in the United States in that they required the boundaries of wetland soils to be identified on maps of individual parcels of property. Doing this required developing new techniques for identifying hydric soils on-site.

The objective of the Clean Water Act was to "maintain and restore the chemical, physical, and biological integrity of the waters of the United States" (ibid). To accomplish this, Section 404 of the Clean Water Act essentially gave the U.S. Army Corps of Engineers the authority to issue permits for (i.e., give permission for) the discharge of dredged or fill material into waters of the United States, including wetlands.

The Food Security Act contained the "swampbuster" provision that denied USDA program benefits to producers who converted wetlands into cropland after December 23, 1985. Farmers who drained wetlands became ineligible for a number of USDA benefits that included price-support loans, purchases, and payments; farm storage facility loans; federal crop insurance; and disaster payments among other benefits.

In the 1980s, some state legislatures also enacted wetland protection laws. For example, Florida's *Warren Henderson Wetlands Act* of 1984 required that wetlands be identified using both hydric soils and hydrophytic vegetation (Hurt and Brown 1995). The act also required that the edge of the wetlands be placed essentially at the boundary of the hydric soils with the uplands. To enforce such a requirement, field methods for identifying hydric soils needed to be identified. Soil scientists with the USDA SCS in Florida and other state agencies tested the use of the SCS's hydric soil criteria for field identification of hydric soils in wetlands across 27 counties in Florida from 1985 to 1986 (Hurt and Puckett 1992).

Results were disappointing in that the USDA criteria were included as hydric soil areas that were not found in wetlands because they lacked hydrophytic vegetation. The reason for this was related to the duration of saturation. Wetlands in Florida were generally saturated within 30 cm of the surface for periods of 30 days or more during the growing season, while the NRCS criteria in use at the time included soils that could be saturated for as short as 1 week (Hurt and Puckett 1992). To enforce the wetland protection laws in Florida, new field indicators were developed for use in that state. The history of the development of the indicators from 1985 to approximately 1994, and a description of the ones developed were described by Hurt and Brown (1995).

Changing the Hydric Soil Definition to Accommodate Field Identification

The federal wetland protection acts created a new use for hydric soils in that the boundaries of wetlands now needed to be identified on-site. In enforcing Section 404 of the Clean Water Act, the Corps of Engineers developed a *Wetlands Protection Manual* that required for jurisdictional wetlands to be identified using a "three-parameter approach" (Environmental Laboratory 1987). Each wetland was expected to meet separate requirements for wetland hydrology, hydric soils, and hydrophytic vegetation.

In 1986 and 1987, the definition of hydric soil was changed twice to accommodate the needs of the Corps of Engineers who were preparing its *Wetlands Delineation Manual* that would illustrate how the wetland hydrology, hydric soils, and hydrophytic plants were to be identified on a site. In 1986, the NTCHS deleted the phrase "in its undrained condition" from the hydric soil definition. The reference to "…that favor the growth and regeneration of hydrophytic vegetation" was still retained. This revision was used in the Corps' *Wetlands Delineation Manual of 1987* (Environmental Laboratory 1987). From a practical standpoint, this change had no impact on the concept of hydric soils used at the time. It was clearly stated in the Corps' delineation manual that a hydric soil can be both drained and undrained. Furthermore, the manual points out that a drained hydric soil "…may not continue to support hydrophytic vegetation" (Environmental Laboratory 1987, p. 27). This interpretation contradicted the definition of hydric soil that is included in the same manual, and no explanation for this point of view was contained within the manual. Possibly because of these inconsistencies, this definition of hydric soil was short lived.

The definition of hydric soils was revised again in 1987 by adding "in the upper part" in place of "that favor the growth and regeneration of hydrophytic vegetation" (Table 2.1). This definition was used in the second edition of the list of hydric soils (USDA 1987). The definition of 1987 deleted any reference to vegetation and focused only on development of anaerobic conditions in the zone where most roots were expected to be found. No depth was specified for the upper part and this was intentional. Some NTCHS members felt that if a specific depth was defined, then, wetland delineators might be expected to prove where in the soil the anaerobic conditions actually occurred should they find themselves in a court of law as defendants in a suit brought by property owners (W. H. Patrick, Jr., NTCHS member, personal communication). This would be difficult to do because direct measurements of soil redox potential would need to be made over time.

The Corps of Engineers published its *Wetlands Delineation Manual* in 1987 to enable field personnel to enforce the mandates of the Clean Water Act (Environmental Laboratory 1987). The manual proposed detailed soil characteristics that could be used nationally to

TABLE 2.4

Selected Field Indicators of Hydric Soils Listed in the Corps of Engineers *Wetlands Delineation Manual of 1987*

Nonsandy Soils
Organic soils (Histosols)
Histic epipedon
Sulfidic material
Aquic or peraquic moisture regime
Reducing conditions
Soil colors
1. Gleyed soils
2. Soils with bright mottles and/or low matrix chroma
Soil appearing on hydric soil list
Iron and manganese concretions
Sandy Soils
High organic matter content in the surface horizon
Streaking of subsurface horizons by organic matter
Organic pans

Source: Adapted from Environmental Laboratory. 1987. Corps of Engineers *Wetlands Delineation Manual*. Technical Report Y-87-1. U.S. Army Engineer Waterways Experiment Station, Vicksburg, MS.

identify the boundaries of hydric soils in the field (Table 2.4). The Corps' field indicators shown in Table 2.4 allowed a person to go to a site and find where the hydric soils occurred in the landscape. The hydric soil was expected to meet at least one of the field indicators shown. A soil appearing on the hydric soil list was considered as a field indicator as long as the soil profile description made at the site confirmed that it was one of the soils on the hydric soil list. The Corps of Engineers produced later manuals for wetlands delineation that will not be discussed here because the manual of 1987 ended up being the main one used into the mid-2000s (Mausbach and Parker 2001). To enforce the Food Security Act's requirements, the USDA's soil scientists used the field indicators in the Corps of Engineers *Wetlands Delineation Manual of 1987* to delineate wetland boundaries on agricultural lands from approximately 1987 to 1995.

New Tools for Identifying Hydric Soils Were Developed

The definition of hydric soils was modified again in 1994 by the NTCHS (Table 2.1), and this is the current definition to date. The phrase "formed under conditions of" was added because it was believed that the morphology seen in a hydric soil was produced before the soil's hydrology was modified by drainage, or before the soil was protected by dams and levees. It implied that a soil was a hydric soil if it had the field indicators needed for a hydric soil. Field personnel did not have to document hydrology or concern themselves with hydrologic modifications. Thus, a drained soil will be a hydric soil if it retains a recognized field indicator of a hydric soil.

Field identification of hydric soils became more important than the list of hydric soils, because of the enforcement of federal and state wetland protection laws. The NTCHS began to develop its own set of hydric soil field indicators during this time period to expand on those used by the Corps of Engineers. The Corps were required to use the *Wetlands Delineation Manual of 1987* by the U.S. Congress for its wetland delineations (National Research Council 1995). As a result, any new hydric soil field indicators identified after 1987 could not be officially adopted by the Corps for wetland delineations. This created problems with field identification over time, because the 1987 list of field indicators simply did not identify all hydric soils in the United States.

To address this issue, the NTCHS assembled a comprehensive list of hydric soil field indicators that could be used across the United States. These new field indicators incorporated those already identified in Florida, and added additional indicators using the field methods that were found to be successful in Florida of comparing the soils in natural wetlands with those in adjacent uplands (Hurt and Brown 1995). These new indicators were first published in *Field Indicators of the Hydric Soils of the United States* in 1996 (USDA 1996). Examples of some of the common field indicators are listed in Table 2.5. The USDA field indicators built upon the ones first identified by the Corps in 1987. However, while the Corps were using only the 13 field indicators shown in Table 2.4, the USDA's field indicators included at least 40. In most cases, the USDA field indicators were soil layers that had precisely defined depths, thicknesses, and colors for both the soil matrix and

TABLE 2.5

A List of Selected USDA Hydric Soil Field Indicators Used to Identify Wetlands That Are Protected by Federal and State Laws in the United States

Symbol	Name	Required Soil Material	Brief Description[a]
S1	Sandy mucky mineral	Sands and loamy sands	A mucky modified sandy mineral layer 5 cm or more thick starting within 15 cm of the soil surface. Soil organic C concentrations must be between 5% and 12%.
S7	Dark surface	Sands and loamy sand	A layer 10 cm or more thick starting within the upper 15 cm of the soil surface and with a matrix value of 3 or less and chroma of 1 or less. At least 70% of the visible soil particles must be masked with organic material when viewed through a hand lens. When viewed without a hand lens, the particles appear to be close to 100% masked. The matrix color of the layer directly below the dark layer must have a chroma of 2 or less.
F3	Depleted matrix	Loams or clays	A layer 15 cm or more thick beginning within 25 cm of the soil surface, with 60% of the matrix having a chroma of 2 or less and value of 4 or more. Redox concentrations are required when the layer is in A or E horizons. Concentrations must be distinct or prominent in contrast with an abundance >2%.
F6	Redox dark surface	Loams and clays	A layer that is at least 10 cm thick, is entirely within the upper 30 cm of the mineral soil and has a. Matrix value of <3 and chroma of <1, and 2% or more distinct or prominent redox concentrations occurring as soft masses or pore linings, or b. Matrix value of <3 and chroma of <2, and 5% or more distinct or prominent redox concentrations occurring as soft masses or pore linings

[a] United States Department of Agriculture, Natural Resources Conservation Service. 2010. Field Indicators of Hydric Soils for the United States, Version 7.0. In L. M. Vasilas, G. W. Hurt, and C. V. Noble (Eds.). USDA, NRCS, in cooperation with the National Technical Committee for Hydric Soils.

redoximorphic features (mottles). In addition, some indicators were defined on the basis of organic carbon percentage. In all cases, the field indicators were defined so that they could be identified from a soil profile description, and this left little room for the "best professional judgment."

Not all of the Corps indicators were retained by the USDA, and those excluded were aquic or peraquic moisture regime, reducing soil conditions, soil appearing on the hydric soil list, iron and manganese concretions, and organic pans. The USDA field indicators were intended to be found on-site and do not require additional or long-term measurements. Identifying aquic or peraquic moisture regimes would require water table monitoring over time periods of at least 1 year. This is not practical because such measurements are expensive and time consuming. Identifying reducing conditions can now be done easily with dyes, but at the time the USDA field indicators were defined, the use of the dye was not widespread, and most soil scientists had no experience with it while others did not trust the results when they conflicted with the soil scientist's "best professional judgment." The hydric soil list was also not considered reliable to use for identifying hydric soils on individual sites. This would require classifying the soil properly according to *Soil Taxonomy.*

The Hydric Soil Technical Standard

The USDA's field indicators of hydric soils were a major advance in identifying most hydric soils across the United States. However, some on the NTCHS felt that another method of hydric soil identification was necessary that allowed hydric soils to be identified by direct measurements of water table levels and anaerobic conditions. On-site measurements could be used to evaluate soils that were suspected of being hydric soils but did not meet known field indicators.

In 2003, the NTCHS adopted the Hydric Soil Technical Standard that set the requirements for how hydric soils could be identified using measurements of saturation, anaerobic conditions, and rainfall (NTCHS 2007). The standard is used to determine if a soil is a hydric soil when the soil does not have a recognized hydric soil field indicator. It can also be used to collect data that will be used to identify new field indicators. The standard is met if a soil is saturated and anaerobic for 14 consecutive days during the growing season in a year of normal or below-normal rainfall. The requirements were based on the 1994 definition of a hydric soil shown in Table 2.1.

In 2012, the NTCHS modified the hydric soil criteria that were used in developing a national list of hydric soils, because a data base search was no longer being used to assemble the list (Federal Register 2012). The revised criteria are shown in Table 2.6 and represent a major change from the earlier versions. Water table depth requirements and permeabilities were replaced by requirements that the soils either have hydric soil field indicators or that they meet the definition of hydric soil as shown by their meeting the requirements of the Hydric Soil Technical Standard. These changes were made because by this time, the list was being assembled by soil scientists within the individual states, rather than by a data base search. The state soil scientists were selecting the map units that they knew contained the components that either met hydric soil field indicators, or were likely to meet the Hydric Soil Technical Standard. From a practical standpoint, this means that lists were being developed by soil scientists in a state who identified those soils that met one or more of the *Field Indicators of Hydric Soils in the United States* from a study of soil profile

TABLE 2.6

Modifications Made to the USDA's Hydric Soil Criteria in 2011

1. All Histels except Folistels and Histosols except Folists; or
2. Map unit components in Aquic suborders, great groups, or subgroups, Albolls suborder, Historthels great group, Histoturbels great group, or Andic, Cumulic, Pachic, or Vitrandic subgroups that
   a. Based on the range of characteristics for the soil series, will at least in part meet one or more *Field Indicators of Hydric Soils in the United States*, or
   b. Show evidence that the soil meets the definition of a hydric soil
3. Map unit components that are frequently ponded for a long duration or a very long duration during the growing season that
   a. Based on the range of characteristics for the soil series, will at least in part meet one or more *Field Indicators of Hydric Soils in the United States*, or
   b. Show evidence that the soil meets the definition of a hydric soil; or
4. Map unit components that are frequently flooded for a long duration or a very long duration during the growing season that
   a. Based on the range of characteristics for the soil series, will at least in part meet one or more *Field Indicators of Hydric Soils in the United States*, or
   b. Show evidence that the soils meet the definition of a hydric soil

Source: Adapted from Federal Register. 2012. *Changes in Hydric Soils Database Selection Criteria*. Vol. 77, no. 40. National Archives and Records Administration and U.S. Government Printing Office, Washington, DC.

descriptions. Names of those map unit components that met at least one field indicator were added to the list of hydric soils. In cases where water table data, and anaerobic conditions have been assessed, then, the *Hydric Soils Technical Standard* can help identify those map unit components that are hydric by definition.

Regional Supplements to the Corps' Wetlands Delineation Manual

The Corps' *Wetlands Delineation Manual of 1987* served as the basis for wetland delineation for approximately 20 years. In 2007, the first "regional supplement" to the 1987 manual was published for Alaska (USACOE 2007). This was the first of 10 regional supplements that were developed to essentially update the 1987 Corps manual by replacing sections in that manual that described how hydric soils, wetland hydrology, and hydrophytic vegetation were to be used to identify jurisdictional wetlands. The supplements are regional in that they were designed for selected states or regions in the United States as shown in Table 2.7. From a hydric soils standpoint, the regional supplements are important because they have adopted the USDA's hydric soil field indicators. Thus, once a supplement has been implemented in a region, both the Center of Excellence (COE) and USDA will identify hydric soils using the same set of indicators.

Summary

The concept of what a hydric soil is has changed since the 1970s. The definition of a hydric soil was originally based on the soil properties found in natural wetlands with

TABLE 2.7

A List of the Regional Supplements That Have Been Adopted

Region	Examples of States or Islands Included	Implementation Date
Alaska	AK	2007
Arid West	NV, CA, AR, and UT	2008
Western Mountains, Valleys, and Coast	WA, OR, CA, ID, and MT	2008
Great Plains	ND, SD, NE, KS, OK, and TX	2008
Midwest	IA, MO, IL, IN, and OH	2008
Atlantic and Gulf Coastal Plain	LA, MS, AL, GA, FL, SC, NC, and VA	2009
Caribbean Islands	Puerto Rico	2011
North Central and Northeast	MN and WI	2012
Hawaii and Pacific Islands	HI and Guam	2012
Eastern Mountains and Piedmont	KY, TN, WV, PA, and VA	2012

Note: These will replace the field indicators used in the Corps of Engineers *Wetlands Delineation Manual of 1987*.

well-expressed hydrophytic vegetation. This has evolved to include drained soils that may no longer be in jurisdictional wetlands. The purpose for identifying hydric soils has also evolved. Hydric soils were first identified for the mapping program of NWI. The major reason for identifying hydric soils today is to support enforcement of federal regulations that prohibit the filling or draining of wetlands. Even the ways hydric soils are identified have changed. Originally, these soils were identified using hydric soil criteria to search the USDA's soil database. Today, hydric soils are identified by hydric soil field indicators, and for soils that do not have recognized field indicators, the hydric soils can be identified using measurements of saturation and anaerobic conditions and such measurements enable new field indicators to be identified. The changes seen in the concept of hydric soil were largely driven in response to the needs of federal programs that both mapped wetlands and protected them.

References

Cowardin, L. M., V. Carter, F. C. Golet, and E. T. LaRoe. 1979. *Classification of Wetlands and Deepwater Habitats of the United States*. FWS/OBS-79/31. U.S. Fish and Wildlife Service, Washington, DC.

Daniels, R. B., S. W. Buol, H. J. Kleiss, and C. A. Ditzler. 1999. *Soil Systems in North Carolina*. Soil Science Department, North Carolina State University, Raleigh.

Environmental Laboratory. 1987. *Corps of Engineers Wetlands Delineation Manual*. Technical Report Y-87-1. U.S. Army Engineer Waterways Experiment Station, Vicksburg, MS.

Federal Register. 2012. *Changes in Hydric Soils Database Selection Criteria*. Vol. 77, no. 40. National Archives and Records Administration and U.S. Government Printing Office, Washington, DC.

Hurt, G. W. and R. B. Brown. 1995. Development and application of hydric soil indicators in Florida. *Wetlands* 15(1): 32–40.

Hurt G. W. and W. E. Puckett. 1992. Proposed hydric soil criteria (indicators) and their field identification. In J. M. Kimble (Ed.) *Proceedings of the Eighth International Soil Correlation Meeting (VIII ISCOM): Characteristics, Classification, and Utilization of Wet Soils*. USDA/SCS, Lincoln, NE.

Mausbach, M. J. and W. B. Parker. 2001. Background and history of the concept of hydric soils. In J. L. Richardson, and M. J. Vepraskas (Eds.) *Wetland Soils: Genesis, Hydrology, Landscapes and Classification*. Lewis Publishers, Boca Raton, FL.

National Research Council. 1995. *Wetlands: Characteristics and Boundaries*. National Academy Press, Washington, DC.

National Technical Committee for Hydric Soils. 2007. The Hydric Soil Technical Standard. Hydric Soils Technical Note No. 11, http://www.soils.usda.gov/use/hydric/, ftp://ftp-fc.sc.egov.usda.gov/NSSC/Hydric_Soils/note11.pdf

Smith, G. D. 1986. *The Guy Smith Interviews: Rationale for Concepts in Soil Taxonomy*. Soil Management Support Services, Soil Conservation Service, U.S. Department of Agriculture, Washington, DC.

Soil Survey Staff. 1975. *Soil Taxonomy: A Basic System of Soil Classification for Making and Interpreting Soil Surveys*. *Agricultural Handbook No. 436*. Soil Conservation Service, U.S. Department of Agriculture, U.S. Government Printing Office, Washington, DC.

U.S. Army Corps of Engineers. 2007. In J. S. Wakeley, R. W. Lichvar, and C. V. Noble (Eds.) *Regional Supplement to the Corps of Engineers Wetland Delineation Manual: Alaska Region (Version 2.0)*. ERDC/EL TR-07-24. U.S. Army Engineer Research and Development Center, Vicksburg, MS.

United States Department of Agriculture. 1985. *Hydric Soils of the United States*. USDA, Soil Conservation Service, National Bulletin No. 430-5-9, Washington, DC.

United States Department of Agriculture. 1987. *Hydric Soils of the United States*. USDA, Soil Conservation Service, National Bulletin No. 430-5-9, Washington, DC.

United States Department of Agriculture. 1991. *Hydric Soils of the United States*. USDA, Soil Conservation Service, Miscellaneous Publication No. 1491, Lincoln, NE.

United States Department of Agriculture. 1996. *Hydric Soils of the United States*. USDA, Soil Conservation Service, National Bulletin No. 430-5-9, Washington, DC.

United States Department of Agriculture, Natural Resources Conservation Service. 2010. Field Indicators of Hydric Soils for the United States, Version 7.0. In L. M. Vasilas, G. W. Hurt, and C. V. Noble (Eds.). USDA, NRCS, in cooperation with the National Technical Committee for Hydric Soils.

Vepraskas, M. J. 1994. *Redoximorphic Features for Identifying Aquic Conditions*. Technical Bulletin 301. North Carolina Agricultural Research Service, Raleigh, NC.

3

Hydrology of Wetland and Related Soils

James L. Arndt, Ryan E. Emanuel, and Jimmie L. Richardson

CONTENTS

Introduction

Hydrology is the study of water, including the properties and movement of water on and below the earth's surface (Hornberger et al. 1998). Wetland hydrology considers the spatial and temporal distribution, circulation, and physicochemical characteristics of surface and subsurface water in the wetland and its catchment over time and space. Soils record the long-term spatial and temporal distribution and circulation of water because actions of water on soil parent material result in the formation of distinctive soil morphological characteristics. Soil morphology, as used here, is the field observable characteristics possessed by a soil such as soil texture, soil color, and soil structure, and the types of soil horizons present. These soil morphological characteristics, a subset of which is known as "hydric soil field indicators" (see Chapter 8), are directly related to a specific set of hydrologic parameters. Soil horizons, for instance, are layer-like soil morphological features that often develop in response to water movement (see Chapter 1). The study of wetland soils is, therefore, intimately linked to the study of hydrology because hydrology influences soil genesis and morphology.

Soil and Water

Soil, an admittedly complex material, results from the influence of five soil-forming factors (Jenny 1941): (1) organisms, (2) topography, (3) climate, (4) parent material, and (5) time. These factors affect and are affected by water. For example, the biota growing on and in soils are strongly influenced by water's presence, both directly because organisms require water to live, and indirectly because the amount of soil water influences oxygen availability in the soil matrix. Topography frequently directs and controls the flow of both surface and subsurface water to and from a wetland. Climate influences the amount and timing of water availability. Parent material affects the flow of water because it forms the matrix through which surface water infiltrates and through which groundwater flows. The weathering of parent material is directly influenced by water availability. Lastly, time is required for soil development to happen.

Soil also results from the action of four general soil-forming processes: (1) additions, (2) deletions, (3) transformations, and (4) translocations (Simonson 1959; Figure 3.1). Soil is the

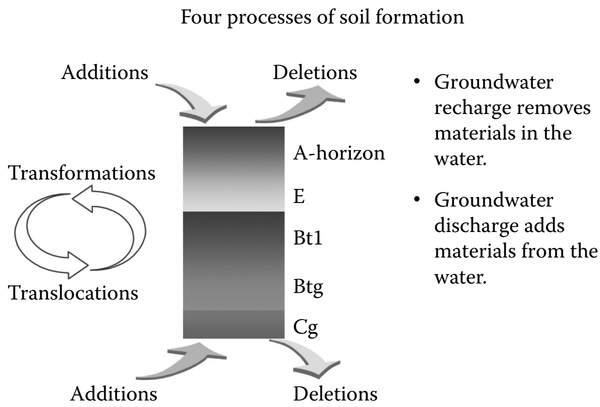


FIGURE 3.1
All four soil-forming processes involve water in some way.

perfect medium in which to study wetland hydrology because all four processes involve water in some way. Water adds material through deposition of eroded sediment and precipitation of dissolved minerals. It transforms soil material through weathering reactions. Water moves (translocates) both solids and dissolved material in mass flow within the soil itself. Water can entirely remove soil material that is dissolved by weathering reactions (transformations), or through erosion of the soil surface.

The study of water and its effects on soil is a unifying principle in soil investigations. The application of hydrologic principles can explain many aspects of hydric soil genesis and morphology that are discussed in detail in other chapters of this book. Similarly, with knowledge of hydrologic principles as a base, the study of hydric soil morphology and genesis relate important information about the nature of wetland hydrology.

Chapter Overview

The study of wetland hydrology requires an introduction to a few basic hydrologic principles. Specifically, hydrodynamics refers to the physical movement of groundwater and surface water to, through, and from a given wetland. Our use of the term hydrodynamics excludes the movement of precipitation and evapotranspiration; however, we are not implying that precipitation, evapotranspiration, and other processes in the hydrologic cycle are irrelevant to an understanding of wetland hydrology. Indeed, the role of the hydrologic cycle in wetland hydrology is discussed further in the next section of this chapter. Most wetlands also exhibit temporal fluctuations in water levels, defined herein as hydroperiod. The water balance of an individual wetland is a fundamental, unique, and distinctive property defined as the budget of water gains, water losses and changes in water storage for a given time period. Water balance is discussed in detail in a later section. Hydrodynamics affect the hydroperiod through controls on the water balance of a wetland. The focus of this chapter will be on hydrodynamics, with a brief discussion of hydroperiod. This discussion is followed by an examination of surface and subsurface water movement. Subsurface water movement is not easily observed and thus requires an introduction to the basic principles of shallow groundwater movement and the influence of both hillslope position and geometry on water movement. Other selected physical aspects of wetland hydrology will be discussed next, followed by a discussion of unsaturated flow and the importance of hydrodynamics at the edges of wetlands. Finally, we will describe the relationship between a hydrology–climatic sequence and soil morphology.

Review of Basic Hydrologic Principles

The Hydrologic Cycle

The endless circulation of water between solid, liquid, and gaseous forms is called the hydrologic cycle. In order to place hydric soil morphology and genesis in the proper context, it is important to recognize that the hydrologic cycle and its associated processes occur at a multitude of spatial and temporal scales. In the broadest scale, water circulates from the oceans to the atmosphere to the land, then back to the oceans (Figure 3.2).

The oceans are the ultimate source and sink for water at the global scale. Evaporation and condensation are the processes by which water changes state from liquid to gas and gas to liquid. The energy to drive these transformations comes ultimately from the sun; however, the processes are important at any scale from microscopic to global. Atmospheric convection and advection along with surface and subsurface flows serve as transport mechanisms. The atmosphere, rivers, lakes, wetlands, soils, aquifers, glaciers, and adsorption of precipitation to surfaces (interception) serve as temporary storage components of the cycle.

Because transport and change of state processes operate at any scale in the hydrologic cycle, water can cycle many times during its journey to and from the ocean. For example,

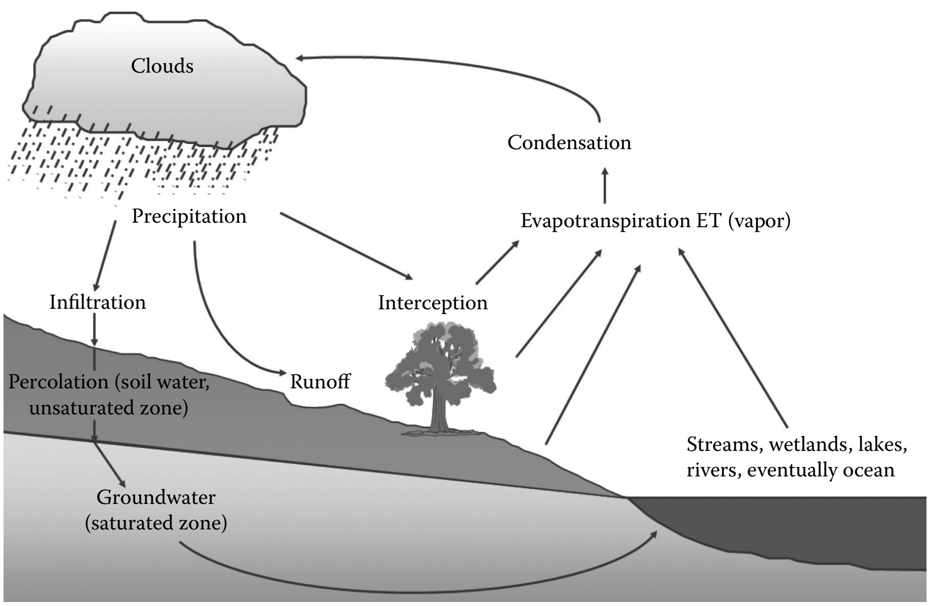


FIGURE 3.2
The global hydrologic cycle depicts various stages of water circulation through the environment. Precipitation strikes the earth where it can be intercepted and evaporated to the atmosphere, infiltrated into the soil, or run off as overland flow.

water vapor in a freezing soil might condense on the surface of a growing ice crystal. When the resulting ice lens eventually melts, the liquid water could move downward into the water table, or it might be taken up by a plant root to be evaporated and released to the atmosphere. In the atmosphere it could condense in a thundercloud and fall as rain onto the surface of a lake, to be stored for days or months prior to evaporation, or it could be released to a stream, with eventual transport to the ocean. In all of its forms, water has a very high capacity to do work. Physical and chemical weathering processes depend on the presence of water.

Basic Water Chemistry, Structure, and Physics

While water is one of the most ubiquitous compounds found in nature, it is also arguably the most unique. A basic review of selected physical properties of water helps in evaluating weathering processes in soils and assessing water movement in saturated and unsaturated soils.

Water consists of two atoms of hydrogen (atomic symbol H) bound to one atom of oxygen (atomic symbol O) (Figure 3.3). The bonds joining the atoms are strongly covalent; thus very large amounts of energy are required to break the bonds holding the water molecule together. The decomposition of water into its constituent atoms rarely occurs, and water molecules are very persistent in nature.

Water is also unusual in that it is found in solid, liquid, and gaseous states within a narrow temperature range that is characteristic of the earth's surface. These characteristics are the direct result of the configuration of the water molecule. The bond formed between the two hydrogen atoms and the oxygen atom is sharply angled at approximately 104.5°, which results in distinct positively and negatively charged regions around the water molecules (Pauling 1970; Figure 3.3). Chemists refer to molecules with distinct positive and negative regions as dipoles. Because water is strongly dipolar, it is strongly attracted to itself (cohesion) and to other charged surfaces (adhesion). An understanding of the cohesive and adhesive properties of water aids in the understanding of the physical state of water in the soil, water movement under saturated and unsaturated conditions, and water's ability to dissolve many substances.

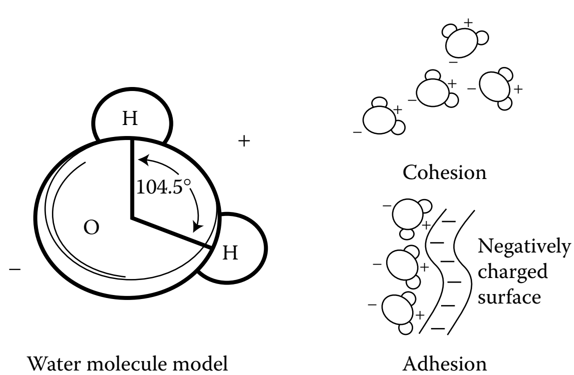


FIGURE 3.3
Structure of the water molecule. Note that the bond angle produces a dipole with opposing positive and negative regions. It is because of the charged dipole that water is attracted to itself (cohesion) and to other charged surfaces (adhesion).

Water the "Universal Solvent"

On a simple level, chemists identify molecules by bond type. Covalent bonds involve electron sharing and are very strong. Ionic bonding involves electron transfers that result in much weaker bonds. Most minerals exhibit mixed bond types that are partly covalent and partly ionic. Molecules with purely ionic bonds are very soluble in dipolar liquids (solvents) such as water because the charged solvent molecules compete with the other atoms in a mineral solid for the bond. Once an atom or a charged portion of the ionic solid is removed from the mineral, the charged molecules of the solvent surround the ion and prevent it from bonding with a solid. Thus, common table salt (sodium chloride, NaCl), a mineral dominantly ionic in character, is much more soluble in water, a dipolar solvent, than in alcohol, which is not as strongly dipolar. Because of its ubiquitous presence and strongly dipolar nature, water is known as a "universal solvent" and is implicated in most, if not all, chemical weathering processes involving geologic and soil materials (Carroll 1970).

Gas Relationships: Aerobic and Anaerobic Conditions

The soil air component of an aerated soil consists of the same N_2, O_2, CO_2, and trace gasses as the atmosphere. The proportions of these gasses change, however, in the soil air. The change is in response to soil biota respiration, which consumes oxygen and organic substrates while releasing carbon dioxide. For example, CO_2 concentrations in soils may exceed ambient atmospheric concentrations by 10× or more, reaching 5000 ppm or higher in aerated soils with active respiration (Riveros-Iregui et al. 2007). Nevertheless, oxygen is replenished to the aerated soil, and carbon dioxide rapidly diffuses to the surface such that soil microbial and plant root respiration is not inhibited. Diffusion of gasses through water, however, is approximately 10,000 times slower than diffusion through air (Greenwood 1961). When water saturates an aerated soil, oxygen diffusion through the water is insufficient to maintain aerobic respiration, and aerobes die or become dormant (Gambrel and Patrick 1978; Skopp et al. 1990). In order to survive under saturated conditions in the soil, organisms evolved adaptive processes to circumvent the lack of oxygen (anaerobic processes). The intensity and duration of these processes are controlled by the amount and persistence of water saturation in the soil, along with other factors.

Basic Hydrologic Principles Describing Groundwater Flow

In a very elementary way, the persistence of groundwater saturation causes a hydric soil to form. However, groundwater is a dynamic component of the hydrologic cycle. Groundwater flow strongly influences the intensity and rate of soil chemical and physical processes that leave numerous morphologic indicators in soil (see Chapter 7). Thus, in addition to the presence or absence of a high water table in a soil, knowledge of the direction, magnitude, and rate of groundwater flow is necessary to place the morphological characteristics of hydric soils in the context of a wetland and its landscape. The direction, magnitude, and rate of groundwater flow are functions of the nature of the porous matrix through which the groundwater flows and the energy status of soil water.

Adhesion, Cohesion, and Capillarity

Soils are porous media containing varying proportions of living and dead organic matter; mineral particles of sand, silt, and clay; water and its dissolved constituents; and gasses. Liquid water interacts with soil solids by adsorption processes. A detailed review of these interactions is beyond the scope of this chapter. For our purposes, it is sufficient to say that hydrophilic surfaces attract and are wetted by water, and hydrophobic surfaces repel water and are not wetted by it, at least initially.

The interactions of adhesive and cohesive forces at solid/liquid interfaces can be described by a simple equation that represents equilibrium between these forces. For example, when a drop of water meets a solid surface, a contact angle (γ) is formed that represents equilibrium between the solid/liquid (σ_{sl}), liquid/gas (σ_{lg}), and solid/gas (σ_{sg}) interfacial tensions (Figure 3.4). At equilibrium, the magnitude of γ defines three classes of substances: (1) those that are not wet (hydrophobic, $\gamma \geq 90°$); (2) those that are partially wet (partially hydrophilic, $0 \leq \gamma \leq 90°$); and (3) those that are completely wet (hydrophilic, $\gamma = 0°$).

The preceding discussion of adhesive and cohesive forces can be extended to describe the phenomenon of capillary rise, which is defined as the height to which water in a capillary tube will rise relative to the free water or water table surface (Figure 3.5). At equilibrium, the adhesive and cohesive forces involved with the surface tension (σ) of water exactly balance the weight of the water in the capillary tube. The relationship is described by Equation 3.1 where H_c is the height of rise in the capillary tube; σ is the surface tension of water; γ is the contact angle between the solid and liquid as defined in Figure 3.5; r is the radius of the capillary tube; ρ is the density of water; and g is acceleration due to gravity.

$$H_c = \frac{2\sigma(\cos\gamma)}{r\rho g} \tag{3.1}$$

Equation 3.1 approximates the height of rise (H_c) in capillary tubes and, within limits, can be used to approximate the thickness of the capillary fringe that exists above the water table in soils with low organic matter. When considering a soil profile with a water table at some depth, we can separate the profile into three distinct regions (vadose, capillary fringe, and saturated zones) defined by the physical state of water relative to the soil

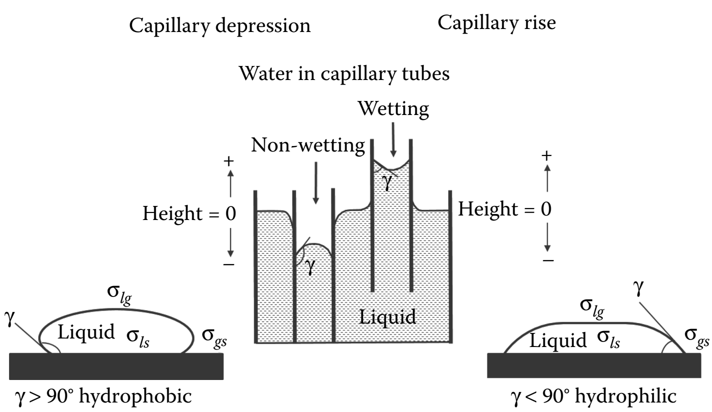


FIGURE 3.4
Contact angle (γ) between a solid and liquid interface determines two classes of substances. Those substances that have $\gamma > 90°$ are not wetted by the liquid and are hydrophobic. Those substances that have $\gamma < 90°$ are wetted by the liquid and are hydrophilic. The upward movement of water ("capillary rise") in capillary pores characterizes hydrophilic solids. Hydrophobic solids exhibit capillary depression. Soils are usually thought of as hydrophilic for water; however, organic matter coatings on soil particles can render them partly to wholly hydrophobic. See Section "Adhesion, Cohesion, and Capillarity" for an explanation of surface tension. (After Kutilek M. and D. R. Nielsen. 1994. *Soil Hydrology*. Catena Verlag, Cremlingen-Destedt, Germany.)

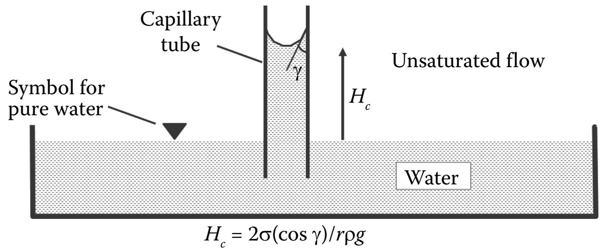


FIGURE 3.5
Height of capillary rise (H_c) relates to the surface tension (σ) of water and air at 20°C. This tension is about 72 D/cm; g is the acceleration due to gravity; and ρ is the density of water. The capillary rise depends on the wetting of soil particles by water and air and the "effective" size of the pores (r) in the soil. Angle γ is the wetting angle between water and the substance. Angle γ is 0° in a fully wetted condition and approaches 90° or a more in repellent condition when no capillary rise occurs (see Figure 3.4).

matrix (Figure 3.6). The water table is defined as the equilibrium level of groundwater in an unlined borehole of sufficient diameter so that capillary rise is negligible. Most of the water found in pore spaces below the water table is "free" water. Free water implies that it is not adsorbed to soil particles.

A capillary fringe of varying thickness exists above the water table (Figure 3.6). While this zone is nearly water-saturated, the water is adsorbed to soil particles to a greater degree than water below the water table. Soil particles with contiguous, very fine pores will have a capillary fringe that rises a considerable height above the water table. Soil particles with pores large enough to drain more easily by gravity will have a capillary fringe that rises to a lower height (U.S. Army COE 1987).

The soil above the water table including the capillary fringe is in the unsaturated or vadose zone. This zone has the potential to contain various amounts of water depending upon the pore size and the height in the soil above the water table. Water in this zone is strongly adsorbed to the soil particles, and many of the air-filled pores are contiguous to

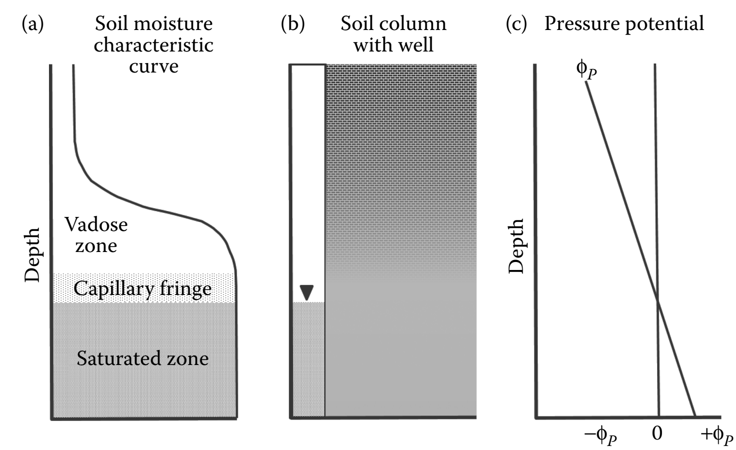


FIGURE 3.6
We can separate the water in a soil profile into three distinct regions: (a) the saturated zone, (b) the capillary fringe, and (c) water in the unsaturated or vadose zone. The pressure potential is positive below the water table and negative above the water table. The capillary fringe is characterized by near saturation with water under negative pressure. The capillary fringe is only a few centimeters thick in most surface soils.

the soil surface and are connected to the atmosphere. At the same time, adsorbed soil water may form contiguous films that partially occupy these pores and may extend from deep in the unsaturated zone all the way to the soil surface, connecting to the atmosphere and promoting soil evaporation (Lehman and Or 2009). The variation of the volumetric water content in the unsaturated zone depends upon the connectivity and size of the interconnected pores, in addition to other abiotic and biotic factors including atmospheric demand for moisture and plant water use.

Implications of the Physical States of Water for Jurisdictional Wetland Determinations

The impact of the capillary fringe thickness on the wetland-hydrology parameter for wetland delineation is not specifically mentioned in the U.S. Army COE (1987) Wetlands Delineation Manual. With regard to a depth requirement for soil saturation in jurisdictional wetlands, the 1987 Manual only states that the wetland hydrology factor is met under conditions where:

> [t]he soil is saturated to the surface at some time during the growing season of the prevalent vegetation. (Paragraph 26.b.3), and
> [T]he depth to saturated soils will always be nearer to the surface due to the capillary fringe. (Paragraph 49.b.2)

Equation 3.1 can be used to calculate the height of capillary rise in soils by assuming constant values for σ, ρ, g, and γ. In pure quartz γ is 0°. Using these constants and expressing length units in centimeters, Equation 3.1 is simplified as

$$H_c = \frac{0.15}{r} \tag{3.2}$$

If we assume that the average effective pore size diameter in medium sands is 0.01 cm, H_c corresponds to 15 cm (6 in.). If we further assume that loams have an average pore size half that of medium sand (0.005 cm), H_c becomes 30 cm (12 in.). Thus, a sandy soil, relatively uncoated with organic matter, with an average effective porosity diameter of 0.01 cm should have a saturated zone extending approximately 15 cm (6 in.) above the free water surface. A loamy soil with an average porosity of 0.005 cm should have a saturated zone extending at least 30 cm (12 in.) above the free water surface (Mausbach 1992). However, these calculations assume that the soil matrix is undisturbed by root channels or other macropores. In reality, macropores commonly complicate the hydraulic properties of soils (Beven and Germann 2013), disrupting the simplified relationship between soil texture and capillary rise shown in Equation 3.2. As a result, the actual height of capillary rise and associated soil saturation can be difficult to determine in many soils.

Various U.S. Army COE district offices (e.g., St. Paul, MN District Office 1996) have provided guidance on the saturation-depth requirement that includes the capillary fringe using Equation 3.2 to compute the height of rise (h). In general, it is assumed that a water table at 6 in. will produce soil saturation to the surface in sandy soils (loamy sands and coarser), and a water table at 12 in. will result in saturation to the surface in loamy, silty, and clayey soils (sandy loam and finer).

An assumption on the thickness of the capillary fringe that is based exclusively on texture, however, is frequently incorrect because the organic matter present in natural soils increases the contact angle (cf. Equation 3.1) and thus reduces the height of capillary rise

(Schwartzendruber et al. 1954; Richardson and Hole 1978). Wetland soils in general, and Histosols or organic soils in particular, have thin capillary fringes due to the presence of large amounts of organic matter that can result in hydrophobic behavior, and strong soil structure that results in a large macropore volume. In many cases water repellency and the corresponding absence of a capillary fringe are observed in soils high in organic matter if the soils are sufficiently dry (Richardson and Hole 1978; National research Council 1995). Soils with even 2% organic matter can have strong structure with large macropores created from fine-textured soils. The aggregates between the pores lack the continuous connection needed for capillarity. The presence of organic matter combined with the confounding effects of soil structure modifying the pore size distribution has been experimentally shown to result in a capillary fringe that is much thinner for the surface layers of most natural soils (Skaggs et al. 1994). Capillarity is normally less than if calculated using only texture due to the complicating effects of soil structure (e.g., root channels, cracks, and other sources of heterogeneity). Many researchers involved in quantification of the soil saturation requirement in jurisdictional wetlands now recommend that the capillary fringe be ignored when evaluating depth to saturation for the surface layers of most natural soils (Skaggs et al. 1994, 1995).

The Committee on Wetlands Characterization (National Research Council 1995) indicated that in the hydrologic assessment of wetlands, the water table depth need not be corrected for a capillary fringe unless field evidence shows that the capillary fringe is large. If the capillary fringe is not substantial, the water table position reasonable approximates the saturated zone for wetland soils and should be the main basis for direct assessment of the hydrology of wetlands (National Research Council 1995).

The Technical Standard for Wetland Hydrology adopted by the COE (U.S. COE 2005) specifically excludes the capillary fringe:

> While its [the capillary fringe] presence has an influence on both plant growth and soil features, the upper limit of the capillary fringe is difficult to measure in the field and impractical as a basis for hydrologic monitoring. The Technical Standard for Wetland Hydrology is based on the depth of the water table because, in most cases, water-table depth can be monitored readily and consistently through the use of shallow wells with either manual or automated data collection. Water-table measurements should not be corrected for a capillary fringe unless other evidence, such as tensiometer readings, laboratory analysis of soil water content, or evidence of soil anoxia, indicates that the height of the saturated capillary fringe is greater than a few inches.

Thus the supplements to the 1987 Manual generally use the position of the water table exclusive of any assumed capillary fringe to be the main assessment of the hydrology of wetlands. For more detailed discussion of the Technical Standard, see Section "Wetland Hydrology and Jurisdictional Wetland Determinations," below.

Energy Potentials and Water Movement

A fundamental principle of fluid mechanics is that liquids flow from areas of high to low potential energy. The total potential energy (Φ_t) of a parcel (or theoretical volume) of water is the sum of various potential energies (potentials), including an osmotic potential (Φ_o), gravitational potential (Φ_g), and pressure potential (Φ_p).

Osmotic potential is the potential energy arising from interactions between the dipolar water molecule and dissolved solids. While Φ_o is important for water flow in plants, it can usually be neglected in soil water flow except in saline soils, which are, by definition, relatively high in dissolved solids. Gravitational potential is the potential energy of position, and can be described by the elevation of a parcel of water above or below some reference datum. Similarly, pressure potential is the potential energy arising from both the pressure exerted by the column of water above the water parcel and the potential energy associated with adsorptive (adhesive) forces between water molecules and soil solids. These two components of Φ_p oppose each other, where the pressure exerted on the parcel by the overlying water column is considered "positive potential," and the pressure due to adsorptive forces is considered "negative potential."

Under saturated conditions, the vast majority of water molecules are far enough removed from solid surfaces that adsorptive forces can be neglected. Φ_p, therefore, is simply due to the pressure of the column of water above the parcel in question. Under these conditions, Φ_p is positive. The water table can be defined in terms of Φ_p as the point where the pressure potential is exactly equal to atmospheric pressure (a condition also known as "zero gauge pressure"). Above the water table, however, there is no column of free water above the zero pressure point except immediately after a rain when water moves downward through the soil in a process called infiltration. After a heavy rain, the larger pores in the soil fill with infiltrating water, moving downward under the force of gravity. At other times, adsorptive forces usually dominate in the unsaturated zone, and as a result Φ_p is negative. Negative pressure potentials (tension) are commonly determined by soil tensiometers. Thus, when one considers a cylinder of soil with a water table at some depth, Φ_p is 0 at the water table, negative above the water table, and positive below the water table (Freeze and Cherry 1979; Hornberger et al. 1998; Heath 2004; cf. Figure 3.6).

Darcy's Law

The first quantitative description of groundwater movement was developed as a result of Henry Darcy's 1856 studies to quantify water flow through sand filters used to treat the water supply for the city of Dijon, France. Darcy's experiment used pressure gauges called manometers to determine the water pressure at varying locations in a cylinder filled with sand, into and out of which there was a constant discharge (Q). The height of water in the manometers relative to a reference level was the "hydraulic head" (H), and the difference in head (dH) between points in the sand divided by the length of the flow path between the points (dL) was the "hydraulic gradient." Darcy then compared Q for different sand textures and hydraulic gradients. He found that the rate of flow was directly and quantitatively related to (1) a factor called the "hydraulic conductivity" (K) that was a function of texture and porosity, and (2) the hydraulic gradient (dH/dL) (Figure 3.7).

Soils and geologic sediments usually form a more heterogeneous matrix for water flow than the sand filters investigated by Darcy. In most situations, the hydraulic conductivity of soils is a function of both soil structure and texture and can be further modified by the presence of large macropores along fractures and root channels. Texture is the relative proportion of sand-, silt-, and clay-size particles. Soil structure is the combination of primary soil particles into secondary units called peds (Brady and Weil 1998). The peds form large pores (macropores) between them which increase the soil's hydraulic conductivity.

The complex spatial distribution of structure and texture combined with the presence of fractures and macropores in natural sediments can confound a Darcian interpretation of groundwater flow unless the characteristics of the flow matrix are taken into account. Laboratory-derived values of hydraulic conductivity are often quite different from field-derived hydraulic conductivity (K) values for the same material (Vepraskas and Williams 1995). Measurements of hydraulic conductivity are scale dependent. The influence of the

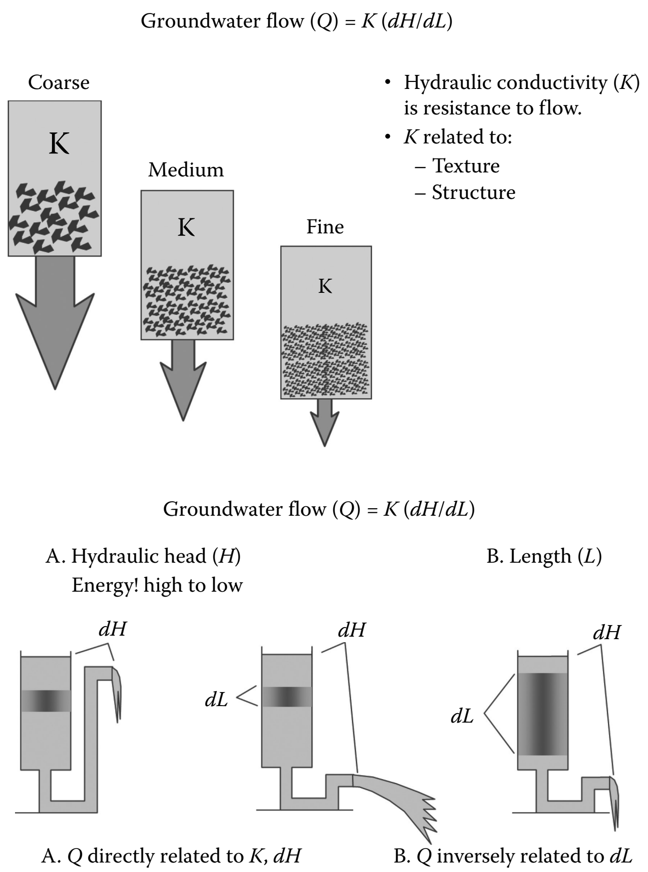


FIGURE 3.7
Saturated flow below the water table can be described by Darcy's Law. The amount of flow is directly related to the hydraulic conductivity (K) and the hydraulic head (dH) and indirectly related to the flowpath length (dL).

nature of the flow matrix on groundwater movement is discussed in detail in section "Soil Hydrologic Cycle and Hydrodynamics."

Assumptions for Darcy's Law

Darcy's law was empirical in nature and was based on experimental observation. Subsequent research has shown that Darcy's law is not valid under conditions where the flow matrix is so fine textured that adsorptive forces become significant (cf. previous section "Adhesion, Cohesion, and Capillarity"), or under conditions where hydraulic gradients are so steep that turbulent flow dominates. However, conditions where Darcy's law does not apply are rarely encountered, and it has become a fundamental tool for quantifying groundwater flow under saturated conditions. Darcy's observations have been validated under most conditions of groundwater flow when the variation of pore size distribution that affects hydraulic characteristics of the flow matrix is accounted for.

It should be emphasized that Darcy's manometers provided quantitative information regarding the total potential of water at the point of interest. In a theoretical exercise, Hubbert (1940) applied equations relating energy and work to prove that the elevations in Darcy's manometers (e.g., hydraulic head) were exactly equal to the total potential energy divided by the acceleration due to gravity. In other words, the elevations in manometers, which are simply monitoring wells (syn. piezometers, see Section on "Methods of Determining the Nature of Groundwater Flow," below), provide quantitative information on energy potentials and energy gradients that can be used in conjunction with information on hydraulic conductivity and flow path geometry to quantify all aspects of groundwater flow at the macroscopic scale.

Methods of Determining the Nature of Groundwater Flow

The concepts of water flow developed above are routinely used to describe groundwater movement in and around wetlands. At a landform or landscape scale, however, it is important to understand how theory interacts with practice for better interpretations of results from groundwater studies. Several readily accessible references are available to understand the principles and methods for quantifying groundwater flow. Good general references include Heath (2004) and Winter et al. (1998). The classic technical reference is Freeze and Cherry (1979). Rosenberry and LaBaugh (2008) provide a good summary of methods for describing groundwater flow and surface/groundwater interactions, and Hornberger et al. (1998) provide a concise description of Darcian groundwater hydraulics and show applications to larger-scale groundwater flow problems.

Piezometers and Water Table Wells

The direction of groundwater flow is determined by monitoring hydraulic head at various locations on the landscape using either piezometers or water table wells (Sprecher 2000). Both devices are similar, commonly consisting of a plastic pipe slotted along some portion of its length and placed in boreholes excavated* below the water table. However, subtle differences between wells and piezometers warrant further discussion.

Piezometers consist of a section of unslotted pipe that is open at both ends or a pipe slotted only at the bottom. The portion of the pipe that is slotted, or the open bottom, may be screened with a "well fabric" to keep soil and sediment out of the pipe while allowing water to flow in. Sand may be packed between the pipe and the borehole wall through the screened zone within the soil profile. Above this sand pack, the remaining area between the pipe and the borehole wall is filled with an impermeable material such as bentonite. When compared to an established reference elevation, the water level in the piezometer represents the hydraulic head at elevation of the slotted interval. This hydraulic head elevation is known as the piezometric surface. It should be emphasized that under conditions of relatively high groundwater flow velocities, the water level in a piezometer may not reflect the piezometric surface under nonflow (static) conditions.

Water table wells, on the other hand, are designed to identify the elevation of the water table ("phreatic") surface (i.e., elevation of the free water surface where the water is at atmospheric pressure). Water table wells most commonly consist of plastic pipe that is slotted to just below the surface or wells slotted at the bottom that have the annular space between the pipe casing and the sides of the borehole filled with coarse sand. The slots and the sand pack act to "short circuit" the piezometric effect or average out the pressure effect. In wetlands, the need to determine the standing water in the upper 15 or 30 cm (sand and other textures, respectively) requires the use of a shallow water table well or several shallow piezometers at a single location.

* Wells can be placed in unlined boreholes, drilled, or driven through the sediments depending on the nature of the sediment and the depth of the well.

Hydraulic heads from at least two piezometers or a water table well are necessary to determine the direction of groundwater flow. Water level elevations from water table wells placed at various points on the landscape can produce a contour map of the water table or piezometric surface that indicates the direction of groundwater flow: water will flow from groundwater mounds (i.e., high head) to groundwater depressions (i.e., low head) along this surface (Noble 2006).

Furthermore, when water table wells are installed at the same location as one or more piezometers (a piezometer nest), the vertical direction of groundwater flow can be determined by comparing water levels in the nested wells. When water levels are the same, stagnant or no flow conditions are indicated (Figure 3.8a). If the water level in the piezometer is lower than that of the water table well, water flow is downward, indicating groundwater recharge (Figure 3.8b). If the reverse is true, then flow is upward, indicating groundwater discharge (Figure 3.8c).

Darcy's law and its mathematical extensions give us the quantitative tools necessary to evaluate groundwater movement in both confined and unconfined, near-surface aquifers. Water table elevations obtained from wells and piezometers indicate local hydraulic heads (H). Local pressure head is the distance between the water table and the screened interval of the piezometer. The distances between wells (L) and water elevations give us the hydraulic gradient in two or three dimensions. Stratigraphy obtained from well logs and actual samples, as well as single-well or multiple-well hydraulic tests, gives us an estimate of hydraulic conductivity within strata. The well and piezometer landscape positions and the magnitude of the water levels reflected in them can be used to relate groundwater recharge and discharge as components of the wetland water balance for a landscape. With

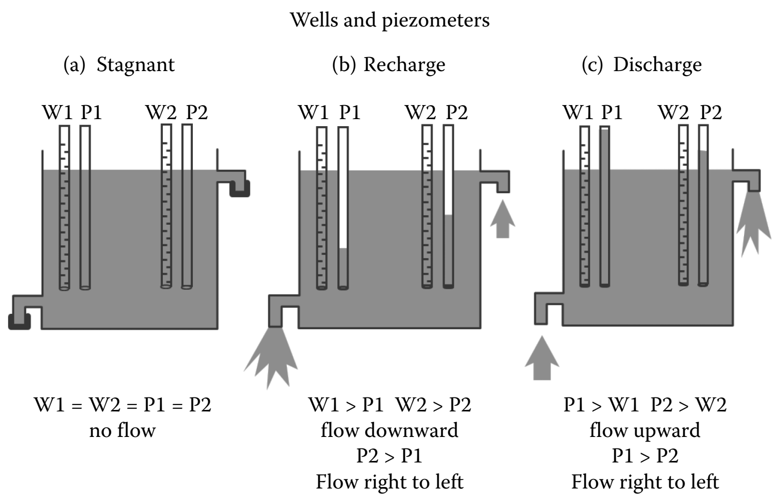


FIGURE 3.8
(a) Stagnant (no flow) conditions illustrated with two sets of wells (W1 and W2) and piezometers (P1 and P2). Piezometers measure the pressure or head of the water at the bottom of the piezometer tube. If the water level of the piezometer is equal to the water level in the well, the hydraulic gradient is 0 and there is no water flow. (b) Recharge conditions illustrated with two sets of wells (W1 and W2) and piezometers (P1 and P2). Piezometers measure the pressure or head of the water at the bottom of the piezometer tube. If the water level of the piezometer is lower than the water level in the well, the hydraulic gradient and water flow are downward. (c) Discharge conditions illustrated with two sets of wells (W1 and W2) and piezometers (P1 and P2). Piezometers measure the pressure or head of the water at the bottom of the piezometer tube. If the water level of the piezometer is higher than the water level in the well, the hydraulic gradient and water flow are upward.

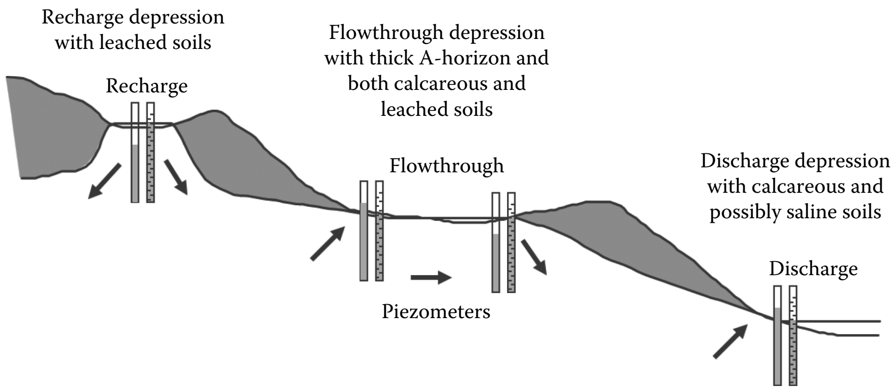


FIGURE 3.9
The magnitude and position of groundwater recharge and discharge as components of the wetland water balance can be identified, and hydric soil morphology can be placed in the context of groundwater flow through the use of Darcy's law combined with well, piezometer, and hydraulic characteristics of the flow matrix.

these data, hydrology can be identified and hydric soil morphology can be placed in the context of groundwater flow on landscapes (Figure 3.9).

Cone of Depression

An analysis of pumping from a well installed below the water table uses the hydrology concepts developed above to demonstrate simply the interaction between saturated flow, the water table, and hydraulic gradient (Figure 3.10). When water is pumped from a well, the water table near the well is depressed as water is removed from the saturated zone and

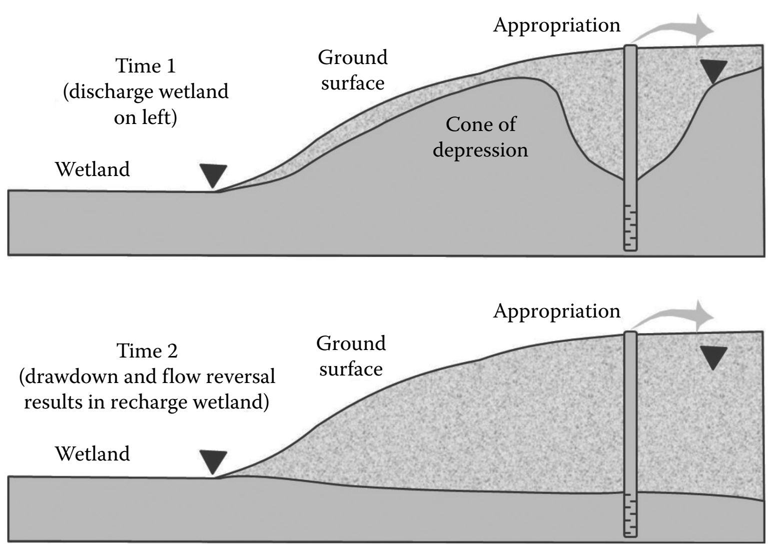


FIGURE 3.10
Domestic water appropriation from a well field in Holland lowered water tables sufficiently to create a groundwater flow reversal in a nearby wetland. (After Schot, P. 1991. *Solute transport by groundwater flow to wetland ecosystems.* PhD thesis, University of Utrecht, Geografisch Instituut Rijksuniversiteirt, 134 p.)

is pumped away. With further pumping, the water table depression progressively moves away from the well, with the water table surface forming the shape of an inverted cone. The shape of the water table depression in the vicinity of the well is appropriately called a cone of depression. The rate of water movement at the water table surface increases with increasing steepness of the water table surface, which represents the hydraulic gradient (dH/dL). As illustrated in Figure 3.10, water will flow faster along the sloping surface of the cone of depression than along the flat surface of the water table away from the cone of depression.

Plants withdrawing water by evapotranspiration produce a drawdown of the water table in a similar fashion, with the effects being more evident at the edge of the wetland where the soil surface is not ponded. Meyboom (1967) showed that phreatophytes (plants capable of transpiring and removing large amounts of water from saturated soil) at the edge of a wetland can change the direction and magnitude of water flow in and around wetlands.

As a broader application, Schot (1991) provided an example of the adverse effects of large-scale domestic groundwater appropriations on adjacent wetlands; these effects may become universal with increasing urbanization. Schot examined the progressive effects of well withdrawals on an adjacent wetland in The Netherlands (a very simplified version is given in Figure 3.10). Prior to and immediately after the initiation of pumping, the wetland received discharge water from the upland. This type of wetland is known as a discharge wetland and would be considered a valuable rich-fen by the Europeans. However, drawdown of the water table by continuous pumping has resulted in a reversal of groundwater flow, such that the wetland now recharges the groundwater (recharge wetland). If pumping were discontinued, the wetland would revert to its natural state as a discharge wetland. If pumping continues, however, the wetland will continue to recharge the groundwater with potentially significant adverse effects to both the water supply and the integrity of the wetland itself. If the wetland water is contaminated, the suitability of the well water may be compromised as the wetland water mixes with the groundwater prior to withdrawal from the well. The wetland's hydrologic regime has changed, and the wetland now loses water to the groundwater instead of gaining water from it. The wetland will certainly get smaller. Depending on the water source, it might dry up altogether. Changes in the water chemistry could also occur because of the removal of the groundwater component to the wetland's water balance. Dissolved solids discharged to the wetland in the groundwater under natural conditions are now removed, and runoff and precipitation low in dissolved solids feed the wetland. The effects of this change dramatically alter the nutrient and plant community dynamics in the wetland, even if it does not desiccate entirely.

Anthropogenic alterations to the groundwater component of wetland hydrology have ramifications for wetland preservation and ecosystem functions and quality. Groundwater pumping in southern Colorado has been linked, for example, to woody shrub encroachment into wetlands of the San Luis Valley, transforming the vegetation structure of these formerly grass-dominated ecosystems and altering their functions (Cooper et al. 2006). A global analysis of internationally important wetlands suggests that regional groundwater withdrawals are altering wetland ecosystems around the world (Verones et al. 2013).

Climate and Weather

The hydrologic cycle and climate are inextricably intertwined. Climate describes the earth's atmosphere for a given place within a specified interval of time, usually decades or longer. Weather, on the other hand, describes the atmosphere for a given place over a short time period. The distinction between weather and climate is important to the study of hydric soils. Hydric soils are assumed to reflect equilibrium between climate and landscape. The transient effects of wet and dry weather will usually not be reflected in hydric soil morphology because the effects of weather occur over too short a period. Weather is reflected in individual or short-term well, piezometer or other hydrological observations, whereas climate variability is often reflected in long-term hydrological records.

Seasonal Observations and Presentation of Precipitation Data

The use of the NRCS WETS tables to interpret the antecedent precipitation "climatic context" for wetland assessment was originally developed to determine the 3-month climatic context for off-site wetland determinations (USDA NRCS 1997). Similarly, the rolling 30-day sum of antecedent precipitation provided in Sprecher and Warne (2000) describe procedures for evaluating whether precipitation prior to a particular date is normal, wet, or dry. Both methods are described in Sprecher and Warne (2000), and are examined in detail along with methods for their calculation in Mohring (2011). Sumner et al. (2009) evaluate both methods along with commonly used modifications of each, and provide detailed descriptions of method implementation and applicability along with recommendations for use.

The NRCS WETS precipitation data have long been used for interpreting wetland signatures on air photos. Sumner et al. (2009) termed this method the "Direct Antecedent Rainfall Evaluation Method" (DAREM). The DAREM method considers precipitation data from the three months prior to the date of interest and weighs those data for length of time since the precipitation contributed to the water budget. Rainfall for the period of the delineation is determined from on-site or web-served sources, and the WETS table that is based on long-term precipitation normals for county-specific precipitation data is consulted to determine dry conditions (3 years in 10 have less than this amount), wet conditions (3 years in 10 have more than this amount), and normal precipitation (precipitation values between dry and wet). A condition value is provided for each month's actual precipitation (3 for wet, 2 for normal, and 1 for dry), and the preceding months are ascribed a weighting value to give greater emphasis to recent when compared to past precipitation (first, second, and third prior months are weighted 3, 2, and 1, respectively). The two values are multiplied together, and the product range indicates the normalcy of the antecedent precipitation: values ranging from 6 to 9, 10 to 14, and 15 to 18 represent dry, normal, and wet antecedent precipitation, respectively. Some of the assumptions that are implicit in the use of this method are:

- Rain was evenly distributed for the month of observation.
- Three months is the proper length of time to evaluate antecedent precipitation even though hydrologic systems vary considerably in their "lag time."
- Snowmelt contributes to wetland hydrology the same as rainfall.

A sample of data output collected from St. Louis County, MN in September is provided in Figure 3.11.

An alternative to the NRCS WETS method is the rolling 30-day total explained in detail in Sprecher and Warne (2000) and Mohring (2011). The method is summarized in Mohring (2011):

1. Obtain daily precipitation data and monthly ranges of normal for your site from the State Climatology office.
2. Calculate in a spreadsheet and plot 30-day rolling totals for the time period of interest.
3. Plot monthly ranges of normal on the plot of 30-day rolling totals.
4. Compare the rolling 30-day sums to the monthly ranges of normal to determine whether antecedent precipitation was within the range of normal.

Precipitation Worksheet Using Gridded Database

Precipitation data for target wetland location:

county: **Saint Louis** — township number: **49N**
township name: **Midway** — range number: **15W**
nearest community: **Morgan Park** — section number: **35**

Aerial photograph or site visit date:
Thursday, October 24, 2013

Score using 1971-2000 normal period

(values are in inches)	first prior month: **September 2013**	second prior month: **August 2013**	third prior month: **July 2013**
estimated precipitation total for this location:	**0.00**	**0.00**	**1.45**
there is a 30% chance this location will have less than:*	2.70	2.94	2.76
there is a 30% chance this location will have more than:*	4.88	4.93	4.78
type of month: dry normal **wet**	dry	dry	dry
monthly score	**3 * 1 = 3**	**2 * 1 = 2**	**1 * 1 = 1**
multi-month score: 6 to 9 (dry) 10 to 14 (normal) 15 to 18 (wet)	6 (Dry)		

Score using 1981-2010 normal period

(values are in inches)	first prior month: **September 2013**	second prior month: **August 2013**	third prior month: **July 2013**
estimated precipitation total for this location:	**0.00**	**0.00**	**1.45**
there is a 30% chance this location will have less than:*	2.86	2.57	2.79
there is a 30% chance this location will have more than:*	4.99	4.41	4.69
type of month: dry normal **wet**	dry	dry	dry
monthly score	**3 * 1 = 3**	**2 * 1 = 2**	**1 * 1 = 1**
multi-month score: 6 to 9 (dry) 10 to 14 (normal) 15 to 18 (wet)	6 (Dry)		

view USDA-NRCS WETS data for Saint Louis County

FIGURE 3.11
Use of the DAREM method to determine the short-term climatic context for a site examined in September 2013, St. Louis County, Minnesota.

The method of rolling 30-day totals has been used to evaluate long-term water table well data (see Figure 3.12 for a graphic illustration of the method).

Sumner et al. (2009) compared the DAREM and the rolling 30-day total method and found the DAREM to be more accurate and preferred to the rolling 30-day precipitation total method. Readers should consult Sumner et al. for a detailed analysis of the suitability and applicability of specific methods to assess precipitation context for wetland delineations and determination of the presence/absence of wetland hydrology.

Establishing the Implications of Longer-Term Perturbations in Climate

The distinction between climate and weather, however, is blurred somewhat during long-term drought and pluvial periods. Climatic interpretations can have serious problems

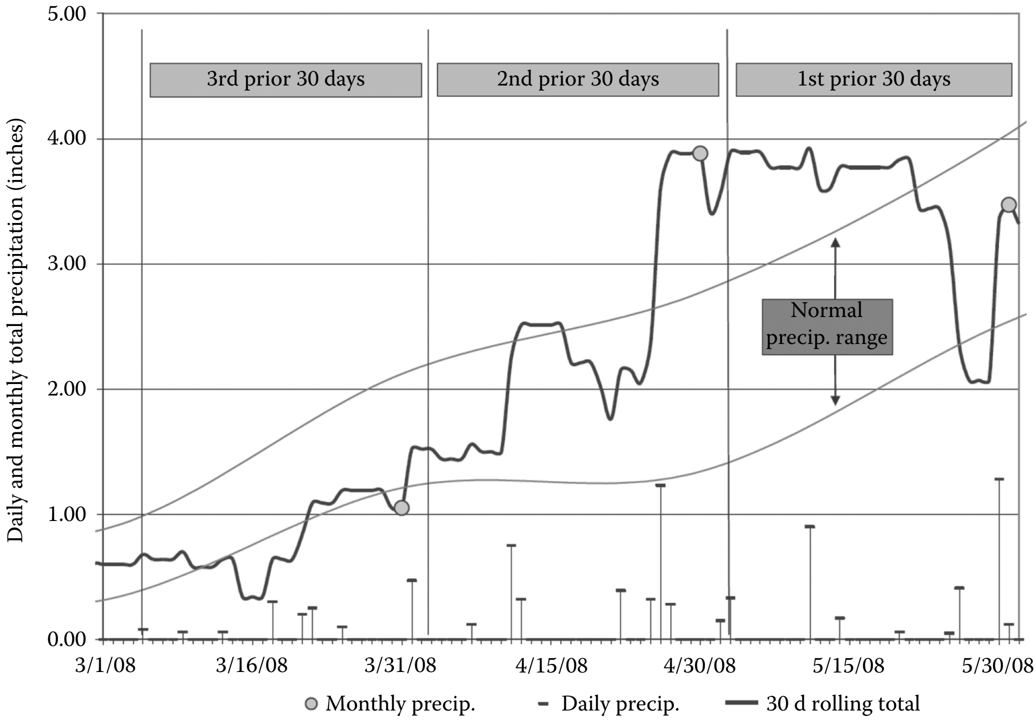


FIGURE 3.12
30-Day Rolling Totals of Precipitation at Cold Spring, MN Overlaid on Graph of Daily Precipitation, Monthly Precipitation, and Range of Normal. The data show that wetland delineations occurring early in the season would have been conducted during a normal precipitation period. However, wetland delineations conducted late April through May would represent wetter than normal conditions.

with regard to regulatory and scientific evaluation. Wetland hydrology during a long-term drought or pluvial period that lasts longer than a decade becomes the "norm" in the minds of people, especially in the case of seasonal wetlands or in wetlands of hydrologically altered areas. Often, relict soil morphology is suspected when it is the morphology that reflects the current local conditions best. The principal difficulty is one of context: is the period in question characteristic of normal conditions or not?

The Palmer Drought Severity Index, developed and used by the National Weather Service, indicates the severity of a given wet or dry period. This index is based on the principles of balance between moisture supply and demand, and it integrates the effects of precipitation and temperature over time. The index generally ranges from −6.0 to +6.0, but as illustrated in Figure 3.13, the index may even reach 8 in some extremes, with negative values denoting dry spells and positive values indicating wet spells. Values from 3 to −3 indicate normal conditions that do not include "severe" conditions. Break points at −0.5, −1.0, −2.0, −3.0, and −4.0 indicate transitions to incipient, mild, moderate, severe, and extreme drought conditions, respectively. The same adjectives are attached to the corresponding positive values to indicate wetter than normal conditions. An example of the Palmer Drought Severity Index applied to the period beginning 1895 and ending 2012 for South Central Wisconsin is shown in Figure 3.13.

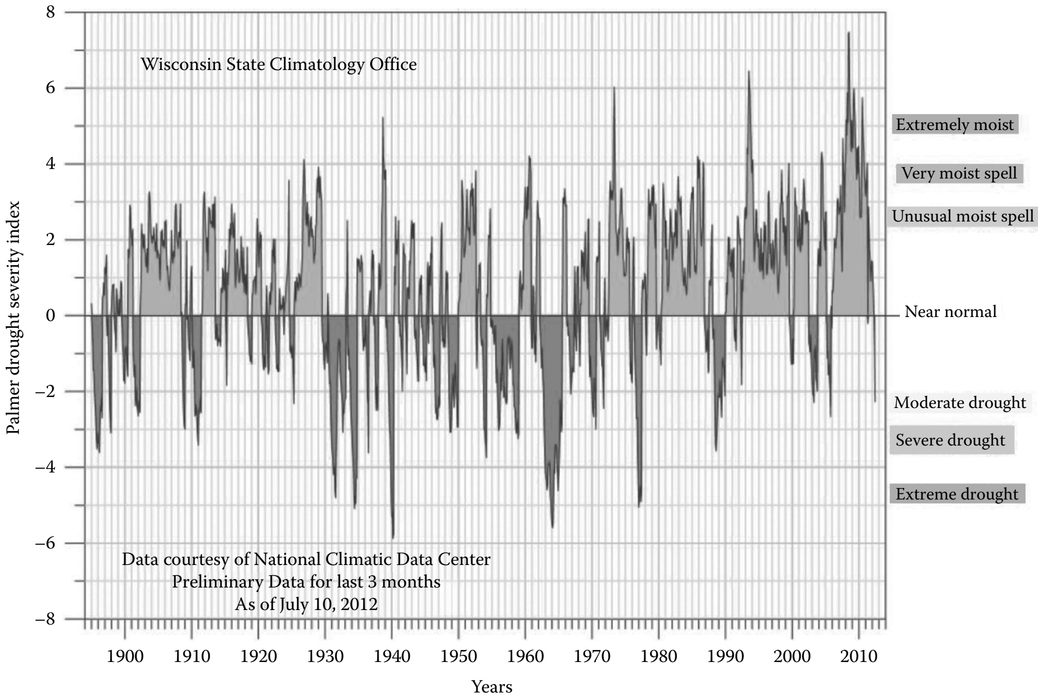


FIGURE 3.13
The Palmer Drought Severity Index (PDI) for South Central Wisconsin for the period January 1895 through June 2012. The data indicate that the period from 1990 through 2012 has been wetter than normal and has been the wettest continuous period since 1905. These data are available on the Internet.

Climate Change

Climate change is becoming a more commonly accepted reality of wetland management (Bates et al. 2008). A large body of literature addresses the potential effects of climate change on many wetland regions, including the Prairie Potholes (Johnson et al. 2005), the metropolitan east coast region (Hartig et al. 2002), the South Atlantic Coastal Plain (Moorhead and Brinson 1995), and the Northwest coast of the United States (Department of Land Conservation and Development 2009). Climate ready communities: A strategy for adapting to impacts of climate change on the Oregon coast (Department of Land Conservation and Development 2009).

With respect to hydrology, climate change could result in several perturbations to the hydrologic cycle with implications for wetlands. Persistent drought would result in reduced input of water and a reduction in wetland area and wetland permanence. Coastal wetlands may become flooded out by sea level rise. Some regions may actually see an increase in wetland areas under more pluvial conditions. Changes in water temperature and persistence can have significant effects on wetland biota ranging from planktonic communities to waterfowl and other macrofauna that are dependent on relatively consistent wetland conditions. With respect to hydrologic impacts, Bates et al. (2008) noted the following:

- Observed warming over several decades has been linked to changes in the large-scale hydrological cycle.

- Climate model simulations for the twenty-first century are consistent in projecting precipitation increases in high latitudes (very likely) and parts of the tropics, and decreases in some subtropical and lower mid-latitude regions (likely).
- By the middle of the twenty-first century, annual average river runoff and water availability are projected to increase as a result of climate change at high latitudes and in some wet tropical areas, and decrease over some dry regions at mid-latitudes and in the dry tropics.
- Increased precipitation intensity and variability are projected to increase the risks of flooding and drought in many areas.
- Water supplies stored in glaciers and snow cover are projected to decline in the course of the century.
- Current water management practices may not be robust enough to cope with the impacts of climate change.
- Climate change challenges the traditional assumption that past hydrological experience provides a good guide to future conditions.

A detailed review of methods to assess the vulnerability of wetlands to climate change is in Gitay et al. (2011). Not only are wetlands vulnerable to climate change, but wetlands have also been recognized for their potential to mitigate climate change. The potential for wetlands to mitigate climate change along with strategies for minimizing climate impacts that may be associated with wetland loss are reviewed in Joosten et al. (2012).

Hydrogeomorphology

Geomorphology is the study of the classification, description, nature, origin, and development of landforms on the earth's surface. Hydrogeomorphology is the study of the interrelationships between landforms and processes involving water. Water erosion and deposition influence the genesis and characteristics of landforms. Conversely, characteristics of the landform influence surface and subsurface water movement in the landscape.

Water Balance and Hydroperiod

The water balance equation describes the water balance in wetlands on the landscape (Figure 3.14). It is deceptively simple, stating that the sum of precipitation, runoff, and groundwater discharge (inputs) are equal in magnitude to the sum of evapotranspiration, surface outflow, and groundwater recharge (outputs), plus or minus a change in groundwater and surface water storage. The process (transpiration) by which plants uptake water and then evaporate some of it through their stomata to the atmosphere, and the process (evaporation) by which water is evaporated directly from the soil or plant surface directly to the atmosphere are combined and called evapotranspiration (ET). Water that infiltrates 30 cm or deeper below the ground surface is usually lost to the atmosphere only through transpiration, with minimal evaporation. Some plants (phreatophytes) draw water directly from the water table. These plants consume large quantities of groundwater and can depress or lower the water table.

When averaged over time, the long-term water balance of an area dictates whether or not a wetland is present. Short-term variations in the water balance of a given wetland produce short-term fluctuations in the water table, defined herein as a wetland's hydroperiod. If inputs exceed outputs, balance is maintained by an increase in storage (i.e., water levels in the wetland rise). If outputs exceed inputs, balance is maintained by a decrease in storage (i.e., water levels in the wetland fall).

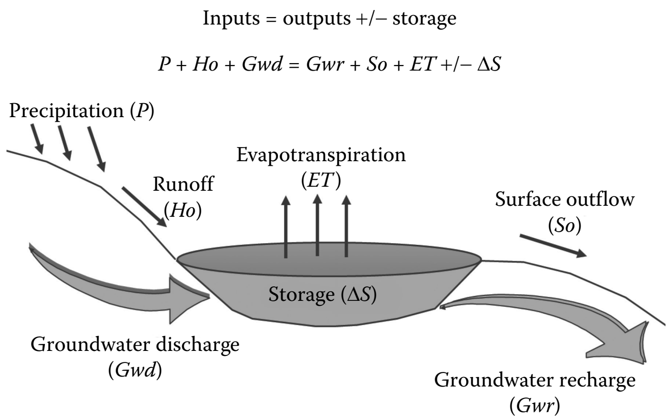


FIGURE 3.14
The hydrologic balance allows for a budget analysis of the water in the environment. By measuring the inputs and outputs along with changes in storage (ΔS), unknown parts of the cycle can be calculated. Various landscapes can be contrasted by knowing a few parameters.

Slope Morphology and Landscape Elements

One of the strongest controls on the water balance of a wetland is topography. Runoff in particular is strongly controlled by topographic factors including slope gradient, slope length, and contributing area (Quinn et al. 1991; Desmet et al. 1999). Other soil conditions being equal (e.g., texture, moisture content, and vegetative cover), for a specific point on a slope:

- Runoff volumes and flow rates will be greater on slopes with higher gradients. Slope gradient influences the speed of runoff and the rate at which runoff infiltrates the soil.
- The greater the runoff volumes will be, the longer the slope above a specific point.
- The larger the catchment area contributing water to a specific point on the slope, the greater the volume of runoff.

Most important for hydric soil genesis is the way in which slopes direct runoff to specific points on the landscape. Wetlands frequently occur at convergent topographic positions on a hillslope that accumulate runoff water. Hydrologists often use combinations of slope gradient, length, contributing area or other topographic variables to identify convergent areas of the landscape where runoff is likely to occur. The topographic index (Bevin and Kirkby 1979) is one such combination of variables that forms the basis for runoff prediction in a common type of topography-driven hydrological model such as TOPMODEL (Beven 1997) that are based on digital elevation data. The topographic Index (*TI*) is obtained from digital elevation (DEM) data as

$$TI = Ln\left(\frac{\alpha}{\tan\beta}\right) \tag{3.3}$$

The contributing area above a specific cell in a digital terrain model is "α" and the "β" represents the down gradient slope.

Landforms consist of slopes having distinctive morphologic elements with widely differing hydraulic characteristics (Figure 3.15). Subsurface water content progressively

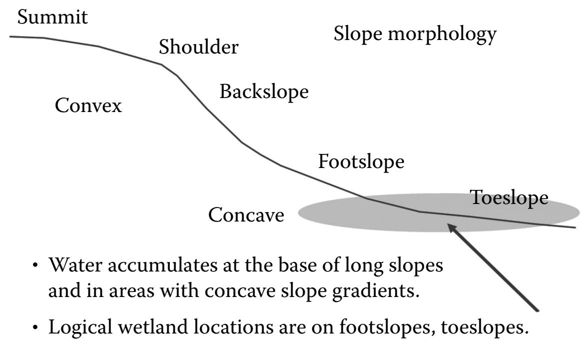


FIGURE 3.15
Hillslope profile position. Wetlands are favored at hillslope profile positions where water volumes are maximized and slope gradients are low. (After Schoeneberger, P. J. et al. 1998. *Field Book for Describing and Sampling Soils*. National Soil Survey Center, Natural Resources Conservation Service, USDA, Lincoln, NE.)

increases downslope as runoff from upslope positions is added to that of downslope positions. A low slope gradient and relatively low soil water content generally characterize the highest (summit) position. Slope gradients increase in the shoulder positions, generally reach a maximum in the backslope positions, and then decrease in the footslope and toeslope (lowest) positions. Footslope and toeslope positions are characterized by maximum water content and minimum gradient. Based on runoff characteristics alone, footslopes and toeslopes in concave positions are logical locations for wetlands because they occur in areas of maximum water accumulation and infiltration.

Slopes exist in more than two dimensions. In three dimensions most slopes can be thought of as variations of divergent and convergent types (Figure 3.16). Divergent slopes (dome-like) disperse runoff across the slope, whereas runoff is collected on convergent (bowl-like) slopes. Plan-view maps of each slope type are shown in Figure 3.16. The presence of convergent and divergent slopes on topographic maps indicates where runoff is focused and recharge is maximized. Convergent areas appear on topographic maps as

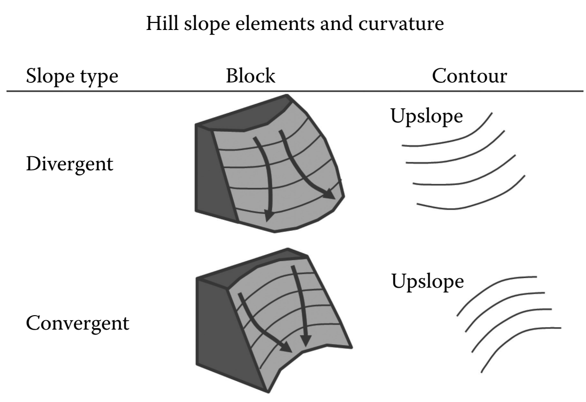


FIGURE 3.16
Hillslope geometry in three dimensions and two directions. Slopes can be thought of as convergent, divergent, and linear (not shown). (After Schoeneberger, P. J. et al. 1998. *Field Book for Describing and Sampling Soils*. National Soil Survey Center, Natural Resources Conservation Service, USDA, Lincoln, NE.)

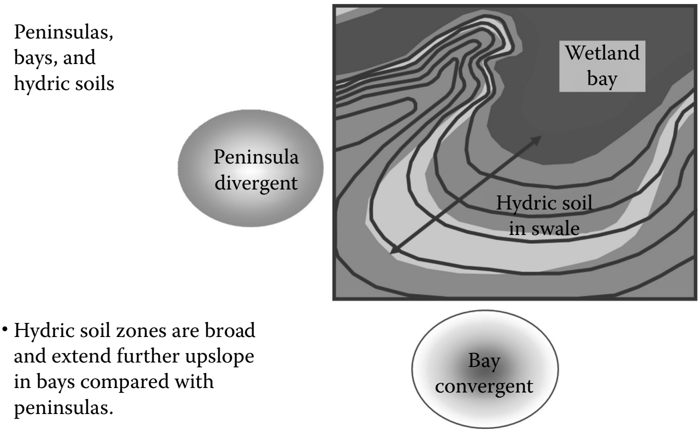


FIGURE 3.17
Swales adjacent to wetland bays are convergent landforms that accumulate water. Divergent water-shedding slopes characterize peninsulas. Hydric soil zones tend to be broad and extend further upslope in bays compared with peninsulas.

depressions in uplands, and peninsulas around wetlands, respectively. Divergent areas appear as knolls in uplands and peninsulas extending into wetlands.

Swales (low depression-like areas) located adjacent to bays in wetlands are in convergent locations, hence, they are characterized by low slope gradients, and they accumulate water. Infiltration and groundwater recharge are maximized, resulting in high water tables. Conversely, peninsulas are divergent landforms often characterized by steeper, water-shedding slopes. The steeper slopes result in both lower infiltration rates and slower groundwater recharge; hence, more precipitation runs off directly to the wetland. Hydric soil zones thus tend to be broad and extend further upslope in bays compared with peninsulas (Figure 3.17). The authors have consistently observed this relationship in the Prairie Pothole Region (PPR) and have frequently used these features for preliminary offsite assessments of wetlands in the region. They can be easily identified on topographic maps and on stereo pair aerial photographs.

The topographic controls on the surface runoff component of the water balance of a given wetland are usually easily understood and directly observable. Topography is also a significant control on the subsurface water-balance components of groundwater recharge and discharge. The relationship, however, is not necessarily direct. Soils and geologic sediments are of equal or greater importance and create situations in which the topographic condition is deceiving because the flow is actually hidden from view in an underground aquifer.

Soils, Water, and Wetlands

The Soil Hydrologic Cycle and Hydrodynamics

The term “wetland” implies wetness (involving hydrology) and land (involving soils and landscapes). Therefore, it is reasonable that an understanding of soil hydrology and soil–landscape relationships is necessary to understand wetland hydrodynamics. The soil

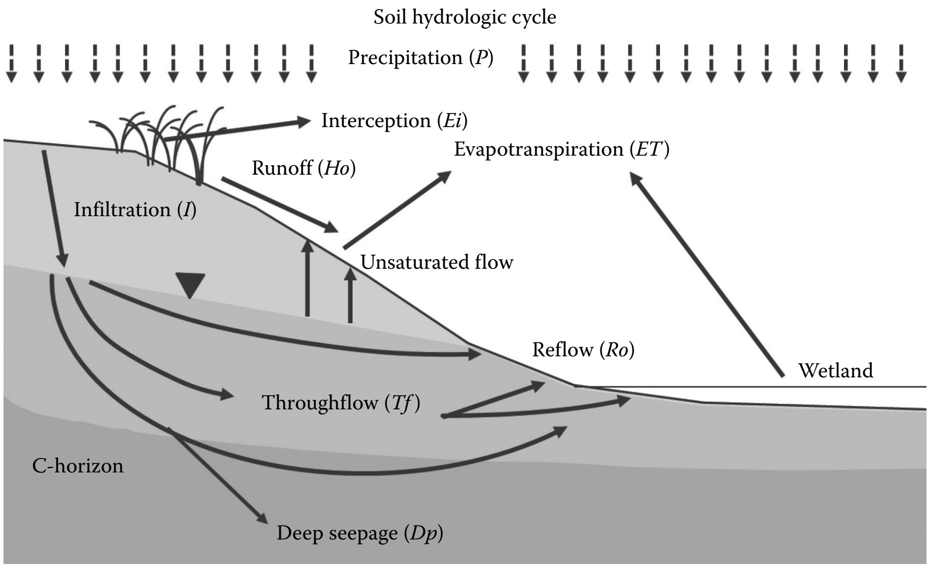


FIGURE 3.18
Soil hydrology includes precipitation, infiltration, surface vegetation interception and evapotranspiration, overland flow, throughflow, deep-water percolation, and groundwater flow. One form of overland flow from a saturated soil is called the reflow. (After Chorley, R. J. 1978. The hillslope hydrological cycle. In M. J. Kirkby (Ed.) *Hillslope Hydrology*. John Wiley & Sons, New York, pp. 1–42.)

hydrologic cycle (Figure 3.18, after Chorley 1978) is a portion of the global hydrologic cycle that includes progressively more detailed examination of water movement on and in the landscape.

Precipitation that falls on the landscape is the ultimate source of water in the soil hydrologic cycle (Figure 3.18). Precipitation water, which has infiltrated, percolates along positive hydraulic gradients until either the gradient decreases to zero, whereupon movement stops and then reverses via unsaturated flow, as water is removed by evapotranspiration, or water movement continues until the wetting front merges with the water table. At this point, groundwater recharge occurs and the water moves by saturated flow in the subsurface. This subsurface, saturated flow usually flows laterally and is called throughflow (*Tf* in Figure 3.18). Groundwater moving by throughflow may discharge at the soil surface and flow as reflow (*Ro* in Figure 3.18). When observed at the soil surface, reflow is often referred to as a seepage face. Deep seepage is the water lost from the local flow system to fracture flow or deeper groundwater that is below the rooting zone of most plants. The amount of water moving as deep seepage is usually less than the amount moving as throughflow.

Landscape-scale or catchment-scale water budget approaches are appropriate for the analysis of wetland hydrodynamics and hydroperiod. The water budget can be expressed by the following budget equation, which is presented graphically in Figure 3.19.

$$P = Ei + Ho + I + \Delta S \tag{3.4}$$

In Equation 3.4, P = precipitation input, Ei = amount of precipitation intercepted and evaporated, Ho = amount of Hortonian overland flow (traditional runoff), I = amount of infiltration, ΔS = change in surface storage. Plants are important in increasing infiltration

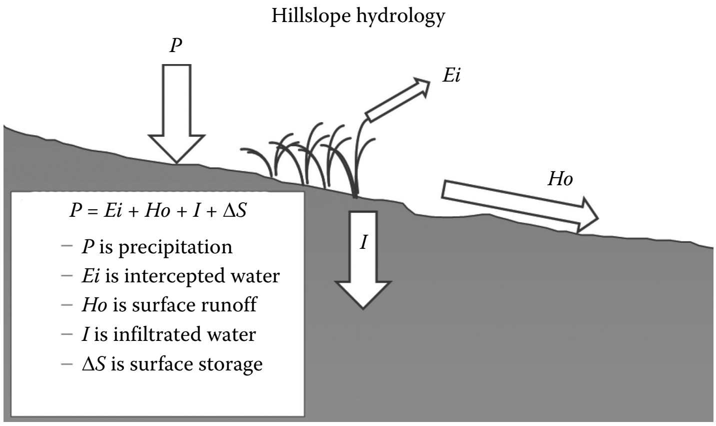


FIGURE 3.19
The surface of a soil separates the water into essentially three parts and two streams. The intercepted water (*Ei*) is sent back to the atmosphere. The water that reaches the surface is split into two flow paths: (1) overland flow (*Ho*) occurs rapidly to the nearby depression, and (2) the infiltrated water (*I*) (groundwater) moves much more slowly along complex paths. Though not readily seen, groundwater can be a very important component of the water balance of many wetlands.

and decreasing runoff and erosion (Bailey and Copeland 1961). Once intercepted by the plant canopy, precipitation may evaporate to the atmosphere or continue flowing to the ground surface as canopy drip or stemflow. Precipitation that is intercepted by the plant canopy loses much of its kinetic energy when it falls or flows to the ground. The reduced kinetic energy results in less detachment and erosion of soil particles at the surface of the soil and less sealing of the pores necessary for water to infiltrate the soil surface.

Water that infiltrates into the soil begins to move downward as a wetting front when the soil surface becomes saturated. Large soil pores, called macropores, transfer water downward via gravity flow. Water that moves through highly conductive macropores can rapidly move past the wetting front (called bypass flow; Bouma 1990). Wetting fronts are frequently associated with the macropores as well; thus, the actual progression of the wetting front in a soil during and immediately after a precipitation event can be very complex.

Soil structure, texture, and biotic activity influence the size and number of macropores, which are most abundant near the soil surface and decrease in abundance with depth. This decrease in number of macropores results in a concomitant progressive decrease in vertical saturated hydraulic conductivity (K_{vs}) with depth in the soil. Horizontal saturated hydraulic conductivity (K_{hs}), however, may remain high across landscapes, reflecting the higher concentrations of macropores in the surface soil horizons.

Transient groundwater flow systems associated with significant precipitation events can impact the hydroperiod of isolated, closed basins, depending on the relative amounts of surface run-on and groundwater flow that are discharged to the pond. The impacts of overland flow on hydroperiod are observed as a rapid rise in pond stage or water table of a given wetland due to the rapid overland flow from the catchment to the pond. The impacts of transient groundwater discharge on pond hydroperiod, however, are not as observable as the impacts of overland flow. The effects can occur over periods of days to weeks depending on the timing, magnitude, and intensity of the precipitation events and catchment geometry.

Shallow but extensive transient, saturated groundwater-flow systems can form in sloping upland soils in the wetland's catchment because of the influence of a permeable surface combined with the presence of a slowly permeable subsoil. Slowly permeable horizons include argillic horizons which have high clay contents, fragipans which are dense layers, cemented horizons such as duripans, as well as frozen soil layers. Lateral groundwater flow through the more permeable surface soil, however, is relatively unrestricted and is driven by a hydraulic gradient produced by the sloping ground surface within the wetland's catchment. The groundwater in this transient groundwater system flows slowly downslope. A portion of groundwater in these transient, shallow flow systems may be discharged to the soil surface upslope of the wetland as reflow, a component of runoff (Figure 3.18). Another portion is discharged to the wetland through seepage at the wetland's edge. A third portion remains as stored moisture when saturated flow ceases. The influence of groundwater discharge on a wetland's hydroperiod (producing a visible water level change) is not immediate because groundwater flow in soil–landscapes is slow relative to surface flow. Significant amounts of water, however, can be discharged to the pond over a period of days or weeks that can maintain the more rapid stage increases produced by surface flow.

The importance of hillslope geometry is illustrated in Figure 3.20. Concave hillslopes, particularly those that are concave in more than one direction, tend to concentrate overland flow, thus maximizing throughflow, interflow, and reflow. During precipitation events, the saturated zone that contributes to reflow increases in area upslope. These saturated areas are potential sites for the genesis of hydric soils.

Water flowing on soil–landscapes can occur as Hortonian overland flow (*Ho*) spawned by precipitation or snow-melt, or it may occur as reflow (*Ro*). Overland flow moves rapidly compared to groundwater. Overland flow contains little dissolved load but carries most of the sediment and usually leaves the sediment on wetland edges or the riparian zone (area along a stream bank) adjacent to stream channels. The magnitude of Hortonian overland flow is inversely proportional to the amount and type of ground cover. Ground cover, moreover, is related to land use.

The water budget for infiltrated water can be expressed by the following equation (after Chorley 1978), which is graphically presented in Figure 3.21:

$$I = Tf + Dp + ET + \Delta SW \tag{3.5}$$

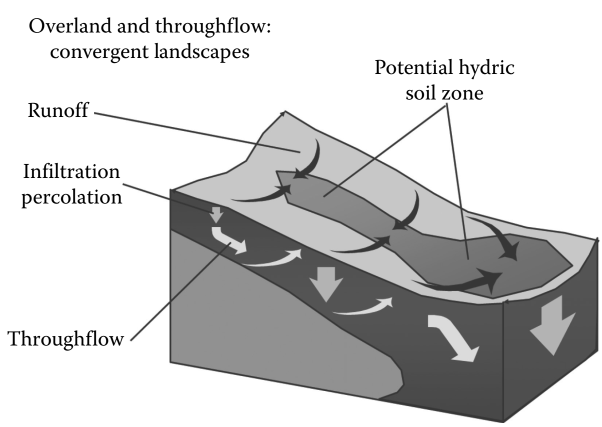


FIGURE 3.20
Illustration of soil hydrology on landscapes with multidirectional concave hillslopes. Water flow converges from the sides as well as from headslope areas. During precipitation events, the saturated zone expands upslope to contribute to increased reflow.

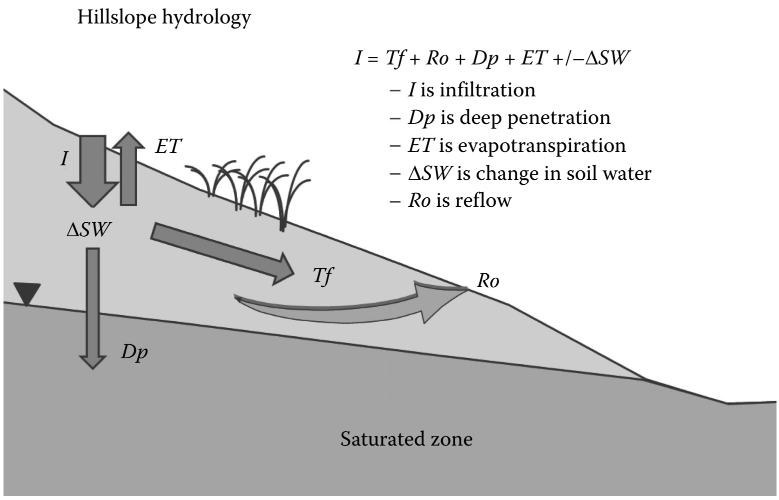


FIGURE 3.21
Water that infiltrates can (1) be used by plants or evaporated, (2) flow downslope in large pores, (3) flow away from the soil surface as deep water penetration, or (4) be added to or removed from the stored soil water. The downslope movement of groundwater (throughflow) discharges at pond edges. Much of the groundwater flows in transient, surficial groundwater flow systems formed in response to significant precipitation events.

where I = infiltration, Tf = throughflow (also called lateral flow or interflow), Dp = deep seepage, ET = evapotranspiration, and ΔSW = change in soil water. The units are usually inches or centimeters of water.

Effects of Erosion, Sedimentation, and Hydroperiod on Wetlands

Land-use changes in a wetland's catchment can alter the wetland's hydrodynamics. Tillage in prairie wetlands, for instance, results in increased runoff and discharge into the wetlands. One of our colleagues working on soils of prairie wetlands relates the story of how his parents had a pair of cinnamon teal ducks nesting in their semipermanent pond in the pasture of their dairy operation. The parents switched from dairy to cropland and plowed the pasture that was the catchment for the pond. The pond became inundated more quickly in the spring; however, it also dried out much sooner and the nesting habitat was lost. The cinnamon teal became a fond memory!

High intensity rains on bare, tilled ground result in high levels of runoff and considerable erosion of the soil that fills depressions with sediment. Runoff and eroded sediments are transported downslope until they are deposited in low-relief areas, including wetlands, and fill the depressions to a degree that they no longer function as wetlands. Conversely, on well-vegetated landscapes more infiltration results in less sediment production. Freeland (1996) and Freeland et al. (1999) observed large amounts of recently deposited sediments as light-colored surface alluvium overlying buried A-horizons in wetlands surrounded by tilled land. No sediments, however, were observed on the soils in wetlands with catchments with native vegetation. Small depressions, in particular, are functionally impacted by even small amounts of sediment. The functions relating to storage of water are particularly disturbed by sediment.

Tischendorf (1968) noted that in 14 months of observation in the southeastern U.S., 55 rainstorms did not produce overland flow in the upper reaches of their forested watershed in Georgia, although 19 storms had enough intensity to produce runoff hydrographs. Flood peaks were related to saturated areas near streams. These areas enlarged during the storm event due to throughflow (interflow), and the associated reflow contributed to overland flow. Kirkham (1947) observed that with intense precipitation, the hilltops had vertical downward flow (recharge), the middle slopes were characterized by throughflow, and the base of slopes had upward flow or artesian discharge flow. Richardson et al. (1994) observed such flows after heavy rains around wetlands in the Prairie Pothole Region. Runoff, however, is not common on the ground surface of forests or grasslands with good vegetative cover, primarily because of the associated high infiltration rates (Kirkby and Chorley 1967; Hewlett and Nutter 1970; Chorley 1978; Kramer et al. 1992; Gilley et al. 1996). The rate of overland flow can be as much as 3 km/h (Hewlett and Nutter 1970). Groundwater flow is orders of magnitude slower than surface flow. For instance, groundwater flowing through coarse-textured sediments at 1 m/day is considered rapid (Chorley 1978), yet this flow rate is only 1/72,000 times that seen in typical surface runoff.

Urbanization also decreases infiltration and increases runoff. Retention ponds constructed to store stormwater runoff effectively behave as recharge ponds that hopefully help to recharge groundwater and wetlands. Obviously, wetland depressions have an important function in terms of sediment entrapment and runoff abatement if retention ponds are being engineered for use in urban settings, although some action will be needed periodically to remove the sediment from retention ponds and place it back on the landscape.

Fringing Wetlands and Wave Activity

Fringing wetlands of the Hydrogeomorphic Model Classification system are wetlands that border lakes, bays, and other large bodies of open water. They have an upland side and a side that yields to the open water, and are thus transitional from upland to open water conditions. During pluvial (i.e., wet) climatic cycles, high water may rise over the emergent vegetation in fringing wetlands. Waves striking the shoreline during these times erode the shore and result in the subsequent formation of a distinctive landscape (Figure 3.22) that consists of (i) a wave-cut escarpment, (ii) a wave-cut terrace, and (iii) a wave-built terrace. These geomorphic features all have distinct soil textures and other physicochemical

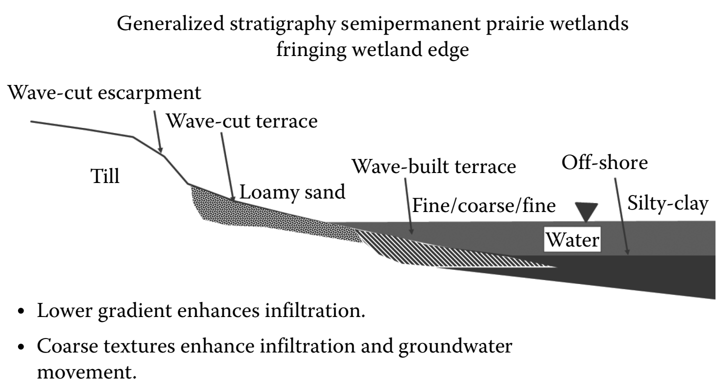


FIGURE 3.22
Fringing wetland edge with an escarpment created by wave erosion that expands the basin width, a wave-cut terrace that is covered with a veneer of gravel, and a wave-built terrace with fine sand and silt. Offshore sediments composed of silts and clays fill the basin and reduce water capacity.

properties. The waves undercut the headlands in steeper areas creating a scarp (an erosional feature). The platform where the waves actually strike is a gently sloping, erosional landform called the wave-cut terrace. While the wave action enlarges the area of the basin, the attendant erosion of the uplands and deposition of the eroded material within the pond decreases overall basin depth and produces a depositional landform called a wave-built terrace that lies pondward of the wave-cut platform.

Although these geomorphic features are not formal indicators of the presence of wetland hydrology in jurisdictional wetlands, wetland scientists performing wetland delineations frequently use these features as secondary indicators of hydrology. These secondary features are incorporated into the "water marks," "drainage patterns," and "sediment deposits" commonly referred to in land ownership disputes around lakes and ponds. We are not referring to "wetland delineation" here but to legal ownership of the land, and such disputes have a far longer history than wetland delineation. Wave-created water-marks around lakes are used to determine public vs. private ownership and access rights of the public around lakes in the Dakotas and Minnesota.

Effects of Saturated and Unsaturated Groundwater Flow on Wetlands

The preceding wave-cut and wave-built landscape is an example of how hydrology and landform interact to produce a distinctive hydrologic pattern in fringing-depressional wetlands. After intense runoff-producing precipitation events, the relatively level sand and gravels on the wave-cut terraces enhance infiltration of the runoff water. Beach sediments act as an aquifer, and the underlying sediments act as an aquitard, resulting in lateral groundwater flow. Once infiltrated, the water rapidly moves laterally along a hydraulic gradient through the coarse-textured beach sediments until it reaches the finer-textured silts and clays characteristic of the wave-built terrace. The silts and clays on the wave-built terrace are lower in hydraulic conductivity. Thus they transmit less water. This results in the development of a transient groundwater mound landward of the interface between the coarse-textured beach sediments and the fine-textured, near-shore depositional sediments deposited pondward from the wave-built terrace (Figure 3.22). This specific type of groundwater/surface water interaction with sediment and landform has been shown to have implications for groundwater discharge, salinization processes, and plant community distribution around Northern Prairie wetlands (Richardson and Bigler 1984; Arndt and Richardson 1989, 1993). These processes may be important hydrologic controls for wetlands outside the Northern Prairie region.

Flownet and Examples of Flownet Applications

Flownets

Darcy's law and its mathematical extensions have been employed in groundwater flow modeling since the mid-1800s. However, the presence of complex stratigraphy and topography, coupled with the need for numerous wells and piezometers necessary to characterize water conditions at a complex landscape scale, have limited the simple application of Darcy's law to small-scale studies or studies that deal with very homogeneous materials.

The influence of stratigraphy and topography on groundwater flow systems was not fully appreciated until the advent of numerical methods and computer programs that accurately model groundwater flow in two and three dimensions. One such method produces a flownet, which consists of a mesh of contoured equipotential lines and flow streamlines.

Equipotential lines connect areas of equal hydraulic head along which no flow occurs. Streamlines indicate the path of groundwater flow and are orthogonal to equipotential lines.

A detailed description of numerical methods and procedures used to develop complex flownets is beyond the scope of this chapter. Detailed descriptions of the methods are in most basic groundwater hydrology texts and papers (e.g., Cedargren 1967; Freeze and Cherry 1979; Mills and Zwarich 1986; Richardson et al. 1992). However, simply put, numerical methods place a two- or three-dimensional rectangular network of grid points over the flow system, and Darcy's equation is applied to develop finite-difference expressions for the flow at each node. Boundary conditions and assumptions, coupled with actual and estimated values of hydrologic parameters at specific nodes, are used to interpolate values for these parameters at the remaining nodes. Seminal research encompassing landscape-scale groundwater modeling that was initiated in the 1960s (Toth 1963; Freeze and Witherspoon 1966, 1967, 1968) has expanded into an explosion of research into virtually all facets of groundwater flow and has resulted in the development of numerous groundwater models.

Figure 3.23 provides the salient characteristics of a flownet simulation using Version 5.2 of the program FLOWNET (Elburg et al. 1990). The figure represents the simple situation of groundwater flows in isotropic, homogeneous media with a water table that linearly declines in elevation from left to right. The height of the bars above the cross-section represents the hydraulic head and is equivalent to the water table elevation. Equipotential lines are dashed, streamlines are dotted, and the large arrow indicates the direction of groundwater movement. By convention, adjacent streamlines form stream tubes through which equal volumes of water flow. Fast groundwater flow is indicated in regions where streamlines are closely spaced. Conversely, slow flow is indicated by widely spaced streamlines.

Effects of Topography (1): Closed Basins, Glaciated Topography

The examples that follow use FLOWNET simulations to illustrate the impacts of topography and stratigraphy on wetland hydrology. Real-world examples from recent soil research are provided to reinforce the concepts present in the simulations.

"Flownets"

$$Q = K \frac{dH}{dL}$$

Darcy's law

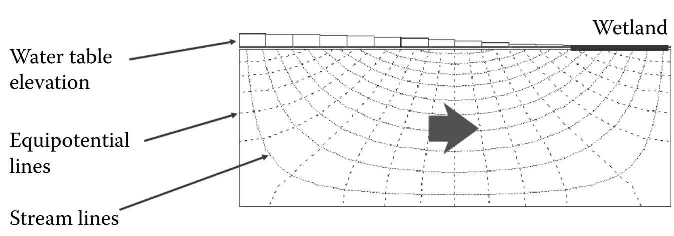


FIGURE 3.23
Arraying equipotential lines (lines of equal hydraulic head) perpendicular to groundwater streamlines creates flownets. Flow is from left to right following equipotential lines. Groundwater recharge occurs to the left of the block diagram, groundwater discharge occurs to the right.

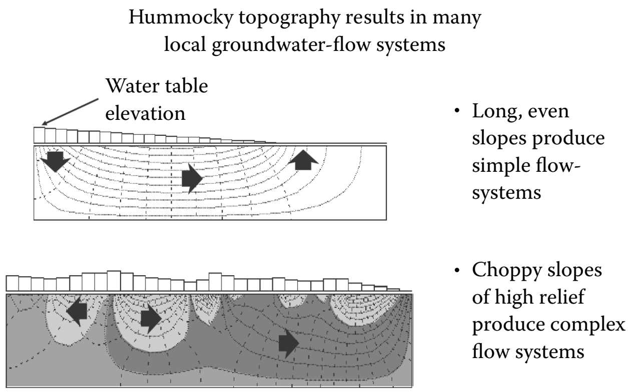


FIGURE 3.24
The upper diagram is a smooth topography with a simple flow pattern. The second indicates the presence of hummocky topography and poorly integrated surface drainage. This creates local flows within larger regional systems. (After Toth, J. 1963. *J. Geophys. Res.* 68: 4197–4213.)

FLOWNET computer modeling accurately simulates or depicts the effect of water table topography on the development of groundwater flow systems as examined in Toth (1963). We assume that the water table topography is a subdued reflection of the surface topography in areas with humid climates. The flownet simulation in Figure 3.24, therefore, illustrates that the presence of a long, regional slope of the water table will result in the development of a simple groundwater flow system. This flow system is characterized by (1) distinct upland recharge zone (upper left portion of the simulation), (2) a distinct zone of throughflow where groundwater is moving approximately horizontally in the middle of the simulation, and (3) a distinct zone of groundwater discharge into a wetland, lake, or river.

The simple flow system described above is in direct contrast to that produced when water table relief is high and complex (Toth 1963). In our FLOWNET simulation, short, choppy slopes that would be characteristic of hummocky glacial topography produce highly complex flow systems consisting of small, locally developed flow systems contained within progressively larger flow systems. The large, bold arrows in Figure 3.24, the second diagram, indicate both localized flow systems that are isolated from each other and the deeper, regional flow system. Groundwater flow within these local flow systems is driven by internal recharge and discharge characteristics. Flow can be with or counter to the regional flow as indicated by the bold arrows. If the water table configuration in Figure 3.24 is persistent, however, there will be no hydrologic groundwater connection between adjacent systems.

The presence of these complex flow systems has a significant impact on the regional hydrogeology. Soluble constituents released by weathering processes that occur during recharge will be transported to groundwater discharge areas. The soluble materials persist within the local discharge system unless removed by some surface transport mechanism, such as wind erosion during drought times or removal in a surface drain in pluvial times. In the Prairie Pothole Region (PPR), where surface drainage is limited or absent, the presence of numerous, hydrologically isolated local groundwater flow systems partly explain why one wetland may be fresh while a neighboring pond is extremely saline.

Effects of Topography (2): Breaks in Slope

Pfannkuch and Winter (1984) observed that breaks in slope, or areas where the slope gradient changes from steep to gentle or flat, were often points of groundwater discharge and

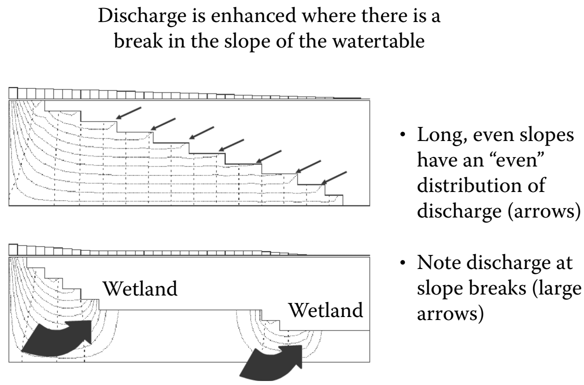


FIGURE 3.25
FLOWNET simulation shows that breaks in slope are frequently groundwater discharge areas occupied by seeps and sloping wetlands.

were frequently occupied by seeps and sloping wetlands. Assuming that the water table is a subdued replica of the land surface, Figure 3.25 shows that their observations are confirmed by a flownet simulation. Water movement within broad, level flats between sloping areas is slow and limited by low hydraulic gradients. Groundwater discharge is focused at the foot of slopes where these hydraulic gradients decrease the greatest amount.

Effects of Topography (3): Wetland Size and Aquifer Thickness

Pfannkuch and Winter (1984) also noted that the intensity of edge-focused groundwater discharge is related to aquifer thickness and wetland size. Because hydraulic head is relatively constant across the ponded wetland surface, the hydraulic gradient decreases rapidly away from the edge. As can be seen in the simulations (Figure 3.26), the effect is magnified when the aquifer is thin and/or the wetland is large. The hydrologic implications are that groundwater discharge is always edge-focused in large ponded wetlands, and that the interior of such large wetlands can be considered to be relatively "stagnant" (or lacking flow) as far as groundwater flow is concerned. This effect is only enhanced

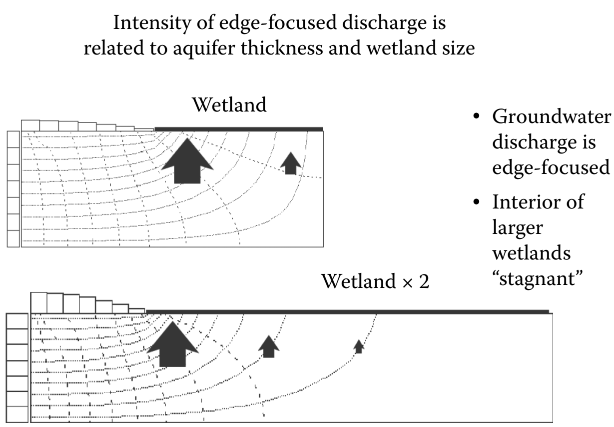


FIGURE 3.26
A FLOWNET illustration of the effect of wetland size and aquifer thickness on groundwater movement. As a wetland increases in size, the tendency is for groundwater to discharge at the wetland edge.

when the wetland edge is also characterized by a break in slope (cf. Figure 3.25 for a simulation). The figure again illustrates the presence of edge-focused discharge and its resulting salinization characteristics.

Effects of Stratigraphy (1): The Effects of Layering

Sediment layering and sediment isotropy/anisotropy are extremely important hydraulic characteristics when considering groundwater flow into and out of wetlands. The FLOWNET simulations discussed above assume topography as the only variable. The flow matrix for these simulations is assumed to be homogeneous, with an isotropic hydraulic conductivity. A sediment layer is isotropic if the hydraulic conductivity within the layer is the same in all directions, and is anisotropic if the hydraulic conductivity differs with direction within the layer. Sediment homogeneity and isotropy are rarely encountered in soil–landscapes. Layering of sediment strata of differing hydraulic conductivity is the usual condition and is caused by the differential action of erosive and depositional processes over time. Most sediments are anisotropic due to depositional and packing processes that favor the lateral orientation of flat, nonspherical particles, and the fact that roots are concentrated near the surface and decrease in abundance with depth. In addition, soil-forming processes create structure and horizons in soils that strongly influence hydraulic conductivity of soils.

In general, lateral groundwater flow is favored over vertical groundwater flow especially in the soil zone, because of (1) the presence of soil horizons and sediment layers of varying hydraulic conductivity, and (2) the presence of anisotropy that favors lateral flow within a given layer (i.e., higher hydraulic conductivity in the horizontal direction). FLOWNET simulations (Figure 3.27) show that layering, in any order, strongly favors lateral flow because of the high flow velocities that are characteristic of the more conductive layer. Given the same hydraulic gradient, flow is much slower in the less conductive layers and is directed primarily downward. The result is that the majority of the flow occurs laterally in the conductive layers. The layer with the lowest hydraulic conductivity limits the speed of downward groundwater flow, and the layer with the highest hydraulic conductivity limits the speed of lateral groundwater flow.

A technique, developed by hydrogeologists, determines the composite horizontal and vertical hydraulic conductivity (K_h and K_v, respectively) for a given stratigraphic section composed of layers of varying hydraulic conductivity (Maasland and Haskew 1957; Freeze

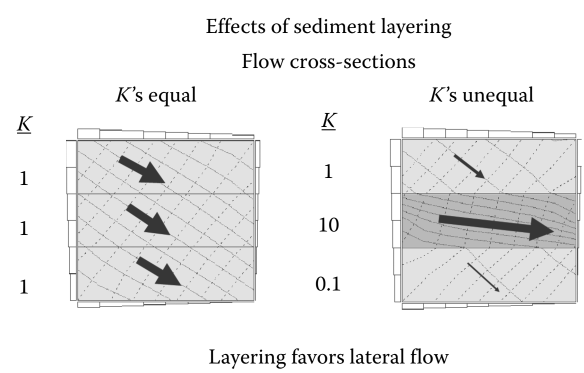


FIGURE 3.27
The effect of layering by soil texture, density or structure creates an increase in lateral flow potential (right-side diagram) when contrasted to the isotropic flow potential (left side) of homogeneous strata.

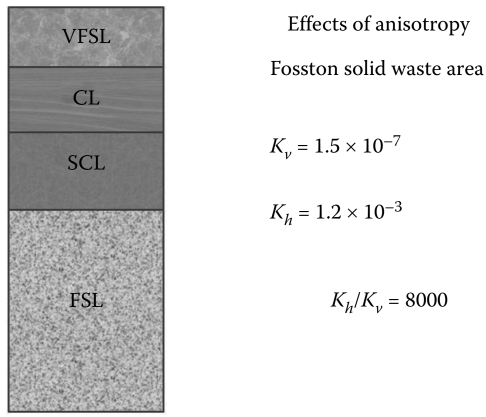


FIGURE 3.28
The concept of anisotropy is that differences between lateral flow and downward flow exist in soils (or rocks). The most restrictive layer (slowest K_v) governs downward movement, and the least restrictive layer (fastest K_h) governs lateral flow.

and Cherry 1979, pp. 32–34). This compositing technique reinforces the significance of the layering impact on groundwater flow. Figure 3.28 provides a situation near a solid waste landfill facility, where the near surface stratigraphy consists of interbedded Pleistocene lacustrine strand and near-shore sediments that vary in texture from clay loam to fine sandy loam. The compositing technique applied to this situation yielded a K_h/K_v ratio of 8000. In other words, for the entire section, groundwater flow was 8000 times faster in the horizontal direction when compared to the vertical direction. In this situation, which contains rather typical sediment layers and hydraulic conductivities, it is obvious that groundwater flow would occur almost entirely within the coarse textured layers and would be lateral in nature. In the field, it is not uncommon for layered heterogeneity to lead to regional composite K_h/K_v values on the order of 100:1 to 1000:1 (Freeze and Cherry 1979).

The impacts of layering are particularly important for transient saturated flow in soils because soils are layered entities that consist of horizons that vary in structure, texture, and hydraulic conductivity. Consider an Alfisol on a slope above a wetland with a well-granulated loamy A horizon, a silty, platy E horizon, and a clay-textured Bt horizon. After a significant precipitation event, water would infiltrate the soil surface and percolate downward; however, the Bt horizon that is low in hydraulic conductivity would limit vertical flow. Throughflow would occur preferentially in the granulated A horizon and the platy E horizon. Groundwater flow would be directed laterally downslope and would resurface as edge-focused discharge at the periphery of the wetland. If rainfall events were frequent enough and of sufficient magnitude, groundwater transferred laterally and downward through soil surface horizons would accumulate on the soil surface at discharge locations and could maintain saturation for a long enough period for hydric soils to develop. This mechanism explains the presence of hydric soils in and adjacent to the bottoms of swales with no evidence of surface inundation, and it also explains the presence of a hydric soil ring above the ponded portions of wetlands.

Effects of Stratigraphy (2): Fine and Coarse Textured Lenses

The presence of soil horizons and sediments with contrasting hydraulic conductivity can have a great impact on both groundwater flow and the resulting presence and hydrologic characteristics of wetlands on the landscape. We can compare groundwater flow

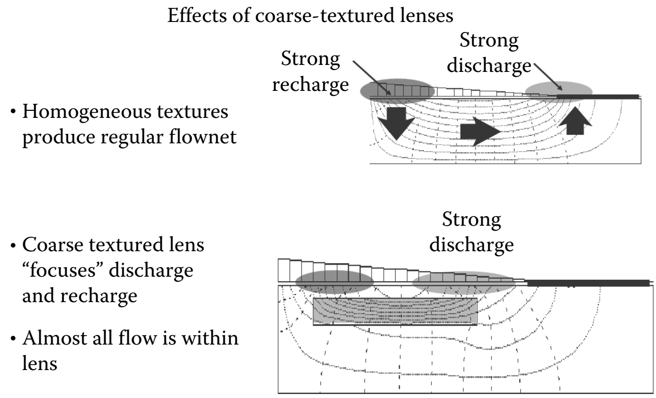


FIGURE 3.29
A comparison of a landscape with homogeneous flow matrix with a similar landscape containing a sand lens embedded in the homogeneous materials. Under saturated flow the sand lens is far more permeable and conductive than the surrounding materials. Water tends to flow into the sand lens and is transported laterally.

in an idealized landscape with a homogeneous flow matrix (cf. Figure 3.23) to a similar landscape containing a sand lens embedded in the homogeneous materials (Figure 3.29). Hydraulic gradients are the same in both illustrations.

The simulation shows that a sand lens acting as a conduit for saturated flow can have a dominant influence on the entire flow system and can strongly influence the hydrologic character of affected wetlands. Under the same hydraulic gradients, flow occurs primarily within the sand lens, with little flow occurring in the fine-textured matrix within which the sand lens is embedded. Groundwater recharge is associated with the up-gradient portion of the sand lens, and groundwater discharge is associated with the down-gradient portion.

Because of much higher hydraulic conductivity, water can be transported laterally in the sand lens, even under small hydraulic gradients. If the sand lens pinches out and terminates, the hydraulic gradient pushes the water to the surface, resulting in a seep. Such seeps can occur even though the sand lens does not crop out at the surface. The effect is exaggerated if the sand lens terminates at the surface, and high volumes of groundwater discharge can form actual spring-heads at these locations. It is important to realize that under these conditions, the sand lens is the flow system. When modeling groundwater flow in such a system, the flow occurring in the fine-textured matrix can be insignificant. Wetlands are frequently formed above these groundwater discharge areas, and many such wetlands have an artesian source of water (Winter 1989).

Areas associated with the up-gradient portion of the sand lens will be strong recharge sites. Soil in these recharge basins will be leached, and often have strongly developed illuvial horizons such as an argillic horizon. Similarly, wetlands associated with down-gradient portions of the sand lens will be strong groundwater discharge sites. Soils in these discharge basins frequently accumulate salts and nutrients and lack leached illuvial horizons. These soils may be organic soils due to the persistent saturation caused by consistent groundwater discharge.

Saline seeps, which are common in the semiarid west, are excellent examples of wet areas resulting from preferential flow in sand lenses and similar zones of higher conductivity. Saline seeps are typically dry for several years in a row because the conductive coarse-textured zones are above the water table which is deeper in dry years. During a pluvial (wet) cycle, however, the water table rises as the sand lens becomes recharged. Once saturated, groundwater flows to points of discharge where the sand lens outcrops or pinches out near the ground surface. The water carries abundant salts that accumulate on the soil surface as discharging groundwater evaporates. Seeps are often discovered during the pluvial cycle by driving a tractor into the seep area, with uncomfortable consequences. Calcareous fens, an unusual type of wetland dominated by groundwater discharge, represent another type of wetland that is commonly associated with coarse-textured lenses embedded in fine-textured sediments.

The presence of less permeable layers in a more permeable groundwater flow matrix also impacts groundwater flow systems and associated wetlands (Figure 3.30). These restrictive layers may have high clay contents, they may contain a restrictive and impermeable soil structure (e.g., platy type), or high bulk densities may characterize them. Groundwater flow in an idealized landscape with a homogeneous flow matrix is compared in Figure 3.30 to a similar landscape containing a less permeable lens embedded in the homogeneous materials. Hydraulic gradients are the same in both cases. The scenario is applicable to any situation where fine-textured sediments underlie coarser-textured sediments, for example, on outwash plains, where fine-textured lacustrine sediments are overlain by coarser outwash sands. In soils, clay-rich argillic horizons frequently have overlying, coarser-textured, and more permeable E horizons that conduct most of the water in sloping landscapes.

The FLOWNET simulation shows that the layer with the lowest hydraulic conductivity restricts downward groundwater flow and forces water to move around it, directing the flow path through more permeable sediments. The result is slower water removal due to shallow gradients that slope to a depression at the edge of the wetland. Additionally, the direct loss of water by ET from the area, poor internal drainage within the overlying sediments, and the potential development of a groundwater mound above the restrictive lens also occur. If the sediments under the restrictive lens are unsaturated, a perched water table results. If the groundwater mound intersects the soil surface, the resulting wetland is similarly a "perched" wetland with soils that have aquic conditions of "episaturation," or water that has accumulated above the soil and tends to move down, or recharge, the groundwater. Soils with episaturation by definition have an unsaturated zone underlying a saturated zone (Soil Survey Staff 2014).

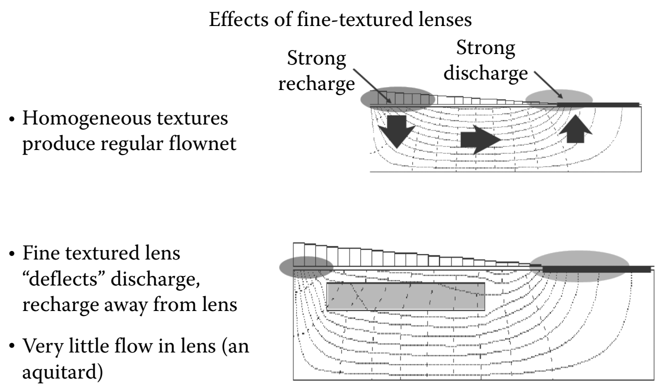


FIGURE 3.30
The rectangle in the FLOWNET is a fine-textured lens that acts to deflect flow around the lens. Flow in the lens or aquitard is nominal. Recharge occurs before the lens or above the lens and flows laterally. Argillic horizons can act like an aquitard on landscapes.

Applications: Wetland Hydrology

Hydrology and Wetland Classifications

Hydrogeomorphic Classification

In order to classify the relationship of landscape and wetlands, we refer to Brinson's (1993) hydrogeomorphic model (HGM). The classes which comprise Brinson's (1993) basic categories in his HGM system separate and group wetlands based on geomorphic setting, dominant source of water, and hydroperiod. These classes reflect wetland processes, such as seasonal depression, because the energy of water is expressed (kinetic energy) or constrained (potential energy) by its soil-geomorphic condition. For example, groundwater in a sloping wetland moves quite differently than groundwater in flats, depressions, fringing, and riverine systems.

A substantial amount of effort over the past several years has resulted in numerous, specific, regionalized HGM guidebooks becoming available. These regional guidebooks provide additional insight into the typical hydrogeomorphology of wetlands in the regions they represent. These regional guidebooks are available on the U.S. Army Corps of Engineers' HGM website.* Depressional wetland systems are the only HGM class covered in the following discussion. The hydrogeomorphic system is discussed in more detail in Chapter 16.

Stewart and Kantrud Depressional Classification

Stewart and Kantrud's (1971) Wetland Classification System, hereafter referred to as the Stewart and Kantrud system, defines hydroperiod for the Northern Prairies of the United States and Canada that were glaciated during previous Ice Ages and retain the imprints of this landscape history. These landscapes are characterized by prairie potholes and other depressional wetlands. The Stewart and Kantrud system divides hydroperiod into three groups based on long-term climatic conditions: (i) normal water levels, (ii) less water than normal, or drought phase, and (iii) more water than normal, or pluvial phase.

The Stewart and Kantrud system uses the definition of hydroperiod to further classify depressional wetlands based on recognizable vegetation zones that develop in response to normal seasonal variations in hydroperiod. They grouped prairie wetland vegetation into zones characterized by (1) distinctive plant community structure and assemblages of plant species, and (2) ponding regime (Table 3.1).

Wetland classes in the Stewart and Kantrud system are based on the type of vegetation zone occupying the pond center, or the wettest part of the pond; thus the wettest zone defines the class. Class II wetlands, for example, are dominated by a wet meadow plant community that experiences only temporary ponding and lacks vegetation typically found in a shallow marsh community. A Class IV wetland characteristically has a central zone dominated by a deep-marsh plant community adapted to semipermanent ponding, and peripheral shallow-marsh, wet meadow, and low-prairie zones, indicating progressively shorter durations of inundation. Figure 3.31 illustrates a "Class IV semipermanent pond or lake" with the relationship of vegetation zones to each other. Concepts of the Stewart and Kantrud system are being extended to classification of nontidal wetlands outside the Northern Prairie region (Brooks et al. 2011).

* http://el.erdc.usace.army.mil/wetlands/guidebooks.cfm accessed March 2014.

TABLE 3.1

Classes and Zones Related to Ponding Regime and Ponding Duration[a]

Class	Central Vegetation Zone	Ponding Regime	Ponding Duration (Normal Conditions)
I	Low prairie[b]	Ephemeral	Few days in spring
II	Wet meadow[c]	Temporary	Weeks in spring; few days after heavy rain
III	Shallow marsh	Seasonal	1–3 months; spring early summer
IV	Deep marsh	Semipermanent	5 months typical
V	Permanent open water	Permanent	Most years except drought
VI	Intermittent alkali	Varies	Varies
VII	Fen	Rarely ponded	Groundwater saturated

[a] Stewart and Kantrud (1971).
[b] The low-prairie zone is too dry to be considered part of a jurisdictional wetland.
[c] The wet meadow zone is the driest part of a jurisdictional wetland.

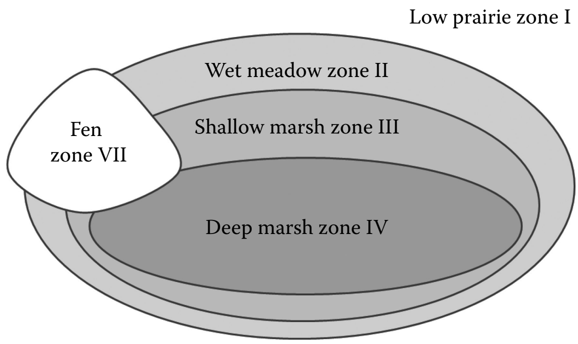


FIGURE 3.31
Arrangement of vegetation zones in a semipermanent pond or lake with a small fen. The wetland edge is the outer wet-meadow or fen zone. The low-prairie is not part of a jurisdictional wetland. (After Stewart, R. E. and H. A. Kantrud. 1971. *Classification of Natural Ponds and Lakes in Glaciated Prairie Region*. U.S. Fish Wildl. Serv., Res. Publ. No. 92. U.S. Gvt. Printing Office, Washington, DC.)

Zonal Classification

The wetland classification system of Cowardin et al. (1979), hereafter referred to as the Cowardin system, is similar in some respects to the Stewart and Kantrud system. The Cowardin system, which is more comprehensive, focuses on vegetation zones rather than on the entire wetland basin. For example, in the Cowardin system, the emergent shallow marsh of Stewart and Kantrud would be separated from the emergent, deep-marsh vegetation zone as a distinct wetland class. Many wetlands characterized under one Stewart and Kantrud class would be characterized under two or more classes in the Cowardin system.

Landscape Hydrology Related to Wetland Morphology and Function

Regional Studies (Macroscale)

Climatology and geomorphology are broad complex disciplines with important applications to understanding hydric soil genesis. Regional wetland characteristics often result from Earth's physical features over broad geographic areas (physiography) interacting with climate differences. For instance, unglaciated areas differ from glaciated areas, and prairie glacial areas differ from forested glaciated areas (Winter and Woo 1990; Winter 1992). Winter and Woo (1990) called divisions at this scale "hydrogeologic physiography" and divided the United States into a few general categories. Climatic criteria, based on gradients between wet/dry and cold/warm extremes, are used by Winter and Woo (1990) to identify a number of varieties of specific regional physiographic types (Figure 3.32). For example, glacial terrains characterized by relatively young landscapes underlain by glacial till with poorly integrated drainage are further broken down by climate into the eastern glacial terrain, which has high precipitation, and prairie glacial terrain (Prairie Pothole Region or PPR), which is characterized by lower precipitation (Figure 3.33). Both regions are fairly representative of a continental climate with cold winter and warm summers. Snow covers the ground 30%–50% of the time. The presence of snow cover and frost during a significant portion of the year has a strong impact on wetlands. Even though

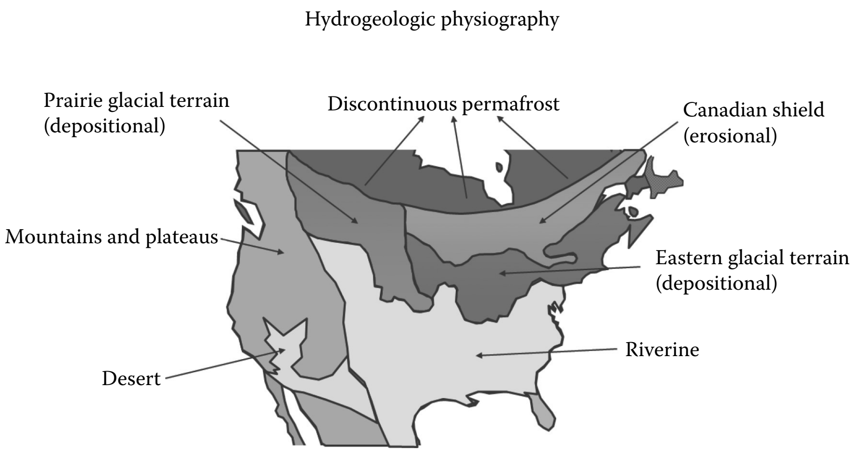


FIGURE 3.32
Climate discriminates the wetlands in the eastern glacial terrain from wetlands in the prairie glacial terrain. (After Winter, T. C. and M.-K. Woo. 1990. Hydrology of lakes and wetlands. In M. G. Wolman, and H. C. Riggs (Eds.) *Surface Water Hydrology. The Geology of North America, v. 0-1.* Geological Society of America, Boulder, CO, pp. 159–187.)

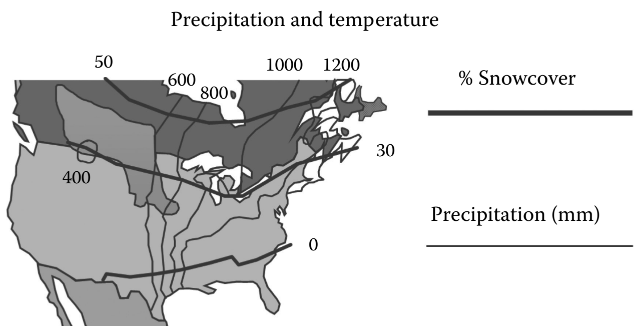


FIGURE 3.33
Contrasting yearly precipitation values in the prairie and eastern glacial terrains. The prairie glacial terrain is added for perspective in relation to the precipitation. (From Winter, T. C. and M.-K. Woo. 1990. Hydrology of lakes and wetlands. In M. G. Wolman, and H. C. Riggs (Eds.) *Surface Water Hydrology. The Geology of North America, v. 0-1.* Geological Society of America, Boulder, CO, pp. 159–187.)

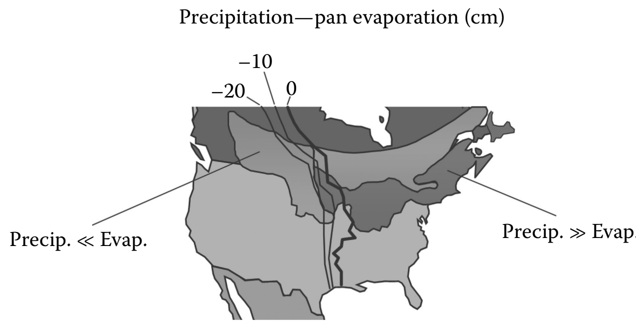


FIGURE 3.34
The border between the prairie and eastern glacial terrains is characterized by the difference between precipitation and pan evapotranspiration. (After Winter, T. C. and M.-K. Woo. 1990. Hydrology of lakes and wetlands. In M. G. Wolman, and H. C. Riggs (Eds.) *Surface Water Hydrology. The Geology of North America, v. 0-1.* Geological Society of America, Boulder, CO, pp. 159–187.)

winter precipitation is usually low, the precipitation that falls is stored in the snow pack, to be released upon spring snowmelt. Because much of the ground is still frozen, runoff is maximized. The period immediately after spring snowmelt is frequently the time of highest water levels for wetlands in these areas, a fact that readily distinguishes cold climate wetlands from those in warmer climates.

It is precipitation, however, that really distinguishes eastern from prairie glacial terrain. The prairie is definitely drier, with average annual precipitation varying from 400 to 600 mm/year compared to the eastern region's 600 to 1400 mm/year.

A more important measure of climate that directly affects wetland hydroperiod, and integrates the effects of temperature and precipitation is the difference between precipitation and pan evapotranspiration. The PPR is characterized by a moisture deficit, whereas the eastern regions have moisture excess (Figure 3.34).

The existence of a moisture deficit in the PPR and a moisture excess in the eastern glaciated terrains has a great bearing on groundwater recharge and discharge relationships. In the eastern glaciated terrain it spawns the development of an integrated surface drainage system. A precipitation surplus is the driving force that causes wetlands to fill to the point where they spill over the lowest portions of their catchments to form these integrated drainage networks. In the eastern glaciated terrain, characterized by moisture, drainage networks are present but poorly integrated due to the youthful, hummocky nature of the unconsolidated tills draped over the underlying bedrock. The PPR landscape is similar geologically; however, low precipitation coupled with moisture deficits ensures that the wetlands usually will not fill to overflowing. The result is a hummocky landscape that is a mosaic of thousands of undrained catchments placed at varying elevations in thick till. Wetlands, varying in ponding duration from ephemeral to permanent, generally occupy highest to lowest positions, respectively, within the catchment.

Groundwater Recharge and Discharge Relationships in Humid, Hummocky Landscapes

Figure 3.35 presents an idealized example of local groundwater relationships in hummocky topography of humid regions characterized by a precipitation surplus. After a precipitation event, a portion of the water falls on the wetland itself (direct interception), a portion is received as runoff from the surrounding catchment, and a portion infiltrates the upland soil and percolates downward or laterally as long as positive hydraulic gradients exist.

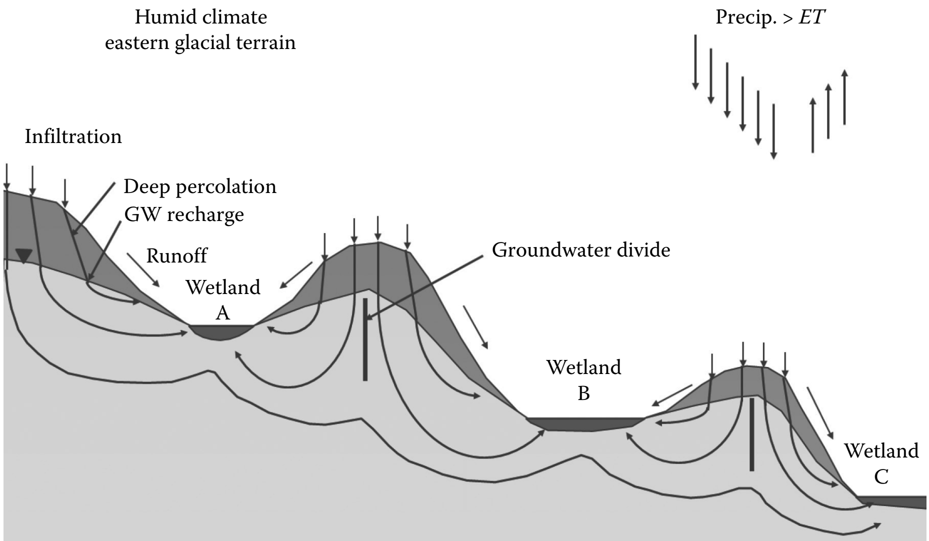


FIGURE 3.35
Humid glacial terrain with groundwater divides in each minor upland. Recharge occurs in uplands, and their soils are leached. Discharge occurs in adjacent wetlands. Surface drainage is developed, although initially it is deranged.

Local groundwater flow systems overlay regional systems. Because precipitation events in the humid region are closely spaced in time, a succession of recharge events drives infiltrated water via deep percolation to the water table. Groundwater is thus recharged in the upland (Figure 3.35), resulting in leached soil profiles. If percolating water reaches the water table faster than it can be discharged to low areas, then a groundwater mound develops under topographic highs. Figure 3.35 represents a generally accepted hydrologic model for groundwater recharge for humid regions. The water table is a subdued replica of the surface topography, and wetlands tend to be foci of local discharge. Groundwater divides form at the crests of the groundwater mounds under topographic highs. These divides are "no-flow" boundaries across which streamlines will not flow; hence, they identify the local flow systems that are superimposed on the regional flow systems in hummocky topography.

Over time, runoff, groundwater discharge, and direct interception will flood the pond until the surface water overtops the lowest portions of the catchment. The resulting meandering, relatively disorganized surface flow (deranged drainage) usually connects wetlands to each other in hummocky eastern glaciated terrain.

To summarize groundwater recharge–discharge relationships in humid regions:

1. Groundwater recharge occurs in uplands, and upland soils are typically leached.
2. Wetlands are "usually" the focus of groundwater discharge.
3. Surface drainages (ephemeral to perennial streams) develop but may be poorly integrated, seeming to meander across the landscape.
4. Many local flow systems overlay regional flow systems.

Groundwater Recharge and Discharge Relationships in Subhumid, Hummocky Landscapes

Figure 3.36 is an example of local groundwater relationships in hummocky topography of subhumid regions that are characterized by a moisture deficit. Wetlands are still

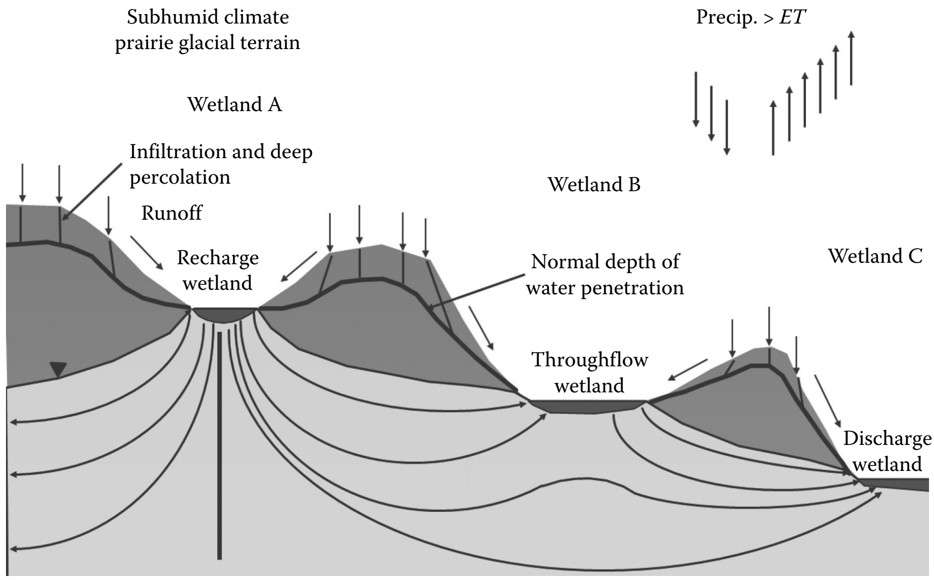


FIGURE 3.36
In subhumid landscapes, the groundwater divide is often in a depression. These landscapes often have flowthrough and discharge wetlands as well as recharge wetlands.

recharged via direct precipitation and overland flow. The longer intervals between precipitation events and the usually intense nature of the events themselves, however, ensure that deep percolation and groundwater recharge does not regularly occur under topographic highs. The groundwater mound is not present under the high because not enough new water infiltrates or penetrates deep enough to reach the water table. Much of the soil water returns to the atmosphere by evapotranspiration before the next recharge event occurs. The overall lack of precipitation coupled with high evapotranspiration further ensures that wetlands will not fill to overflowing.

The above factors result in a landscape dominated by closed catchments and nonexistent surface drainage. Because deep percolation is minimized by the lack of frequent precipitation, interdepressional uplands are relatively uninvolved in transfers of water to and from the water table. In the subhumid PPR, therefore, groundwater recharge and discharge are depression focused (Lissey 1971; Sloan 1972). Seasonally ponded wetlands in upland positions (e.g., Wetland A, Figure 3.36) recharge the groundwater with relatively fresh overland flow and snowmelt. A portion of this recharge water moves downward and laterally into and out of intermediate throughflow wetlands (Wetland B, Figure 3.36), and is subsequently discharged into a low-lying discharge-type wetland (Wetland C, Figure 3.36).

To summarize groundwater recharge–discharge relationships in subhumid regions:

1. Groundwater recharge and discharge are depression-focused.
2. Uplands are relatively uninvolved in groundwater recharge and discharge. Upland soils often contain evidence of limited deep percolation (e.g., presence of Ck horizons, Cky horizons).
3. Surface drainages are limited or nonexistent.
4. Wetlands are distinctly recharge, flowthrough, and discharge with respect to groundwater flow.

A Proposed Wetland–Climatic Sequence

A series of hydrology–climatic sequences was constructed based on experiences in studying soils across climatic regions (Richardson et al. 1992, 1994) and on information from Wetlands of Canada (National Wetlands Working Group 1988). The hydroclimatic sequences were divided into four zones, moving east to west across the northern region of North America: (1) Zone 1—perhumid, (2) Zone 2—humid, (3) Zone 3—subhumid, and (4) Zone 4—semiarid. Zones 1 and 2 relate to the humid region eastern and prairie glacial terrains mentioned in the preceding section. Zones 3 and 4 related to drier terrains.

Excess precipitation in perhumid landscapes leaches the soil of easily soluble materials, including nutrients, and tends to favor acid-forming plants that produce tannin. Tannin is an excellent preservative of organic matter, and that is why it is used to “tan” leather. Tannin restricts bacterial decomposition. The slow loss of mor-type humus or organic material from acid bogs may be largely due to the tannin-created preservation. Mor humus does not mix with the mineral soil nor do bacteria consume it. Its slow decomposition is largely from fungi. Large peatlands, extending for several miles, often cover existing landscapes (Moore and Bellamy 1974). In a depression, organic matter or primary peat accumulates in saturated conditions, reducing the size of the water storage. Next to form are secondary peats that fill the depression up to the limit of water retention. Lastly, acid peats usually formed from sphagnum moss by the growth of “tertiary peat” on the existing peat and often on the land surface around the depression covering the landscape out from the depression (Moore and Bellamy 1974). “Tertiary peats are those which develop above the physical limits of groundwater, the peat itself acting as a reservoir holding a volume of water by capillarity above the level of the main groundwater mass draining through the landscape” (Moore and Bellamy 1974). Such a peat blanket is illustrated in Figure 3.37. Blanket peats are more common in areas of low evapotranspiration and a high amount of precipitation, such as eastern Canada and northern Finland. Water flow is restricted primarily to the peat, and stream initiation is prohibited. In peat basins containing only primary peat, water flow occurs into the basin (cf. humid climatic region). Any water that infiltrates the peat mat and reaches the mineral soil will probably flow laterally below the peat in these landscapes. Secondary peats create a situation that stops or inhibits the

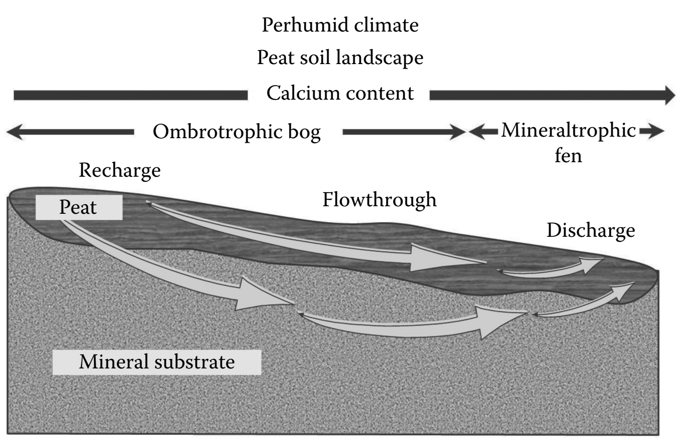


FIGURE 3.37
Perhumid blanket peatland with tertiary peat covering the landscape. Water flows in the peat or in the mineral soil below the peat. Lower areas are enriched with nutrients. Upper areas are distinctly nutrient deficient.

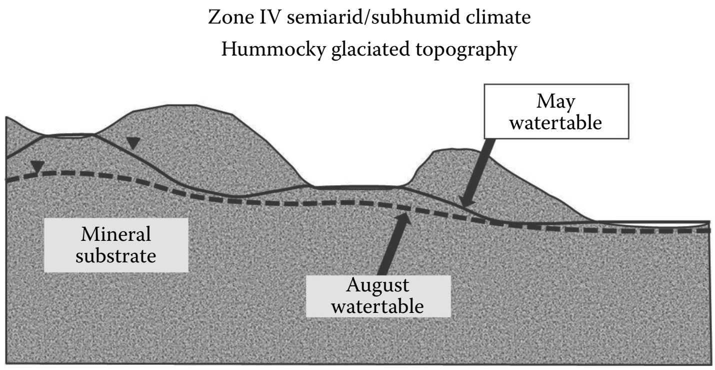


FIGURE 3.38
In semiarid regions with hummocky topography, the depressions are nearly recharge areas. (After Miller, J. J., D. F. Acton, and R. J. St. Arnaud. 1985. *Can. J. Soil Sci*. 65: 293–307.)

growth of stream channels. This lack of channel development results from the fact that water only flows below, on, or in the peat mat. The only water that reaches the peat surface is rainwater and hence is very nutrient poor.

Zone 2 is the same as the humid climate discussed earlier in the section titled Groundwater Recharge and Discharge Relationships in Humid, Hummocky Landscapes, and Zone 3 is the same as the subhumid climate discussed in the section dealing with subhumid, hummocky landscapes. Zone 4 (semiarid) contains dominantly recharge wetlands because the lack of precipitation and high ET precludes the integrated groundwater systems of the aforementioned zones. The climate is so dry that only recharge wetlands or low prairies occur, with a few saline ponds (Figure 3.38). Miller et al. (1985) describe this type of landscape in a semiarid climate. Fifteen of sixteen catchments that they studied were characterized by recharge hydrology and corresponding soil morphologies, such as soils with argillic horizons in the wetlands. Wetland soils were leached, and the surrounding wetland edge soils were calcareous and dominated by evaporites. Many of these soils contained natric horizons.

Generalized Landscapes with Soils and Hydrology

Winter (1988) related two generalized landscapes in an effort to unify the hydrodynamics of nontidal wetlands. The following demonstrates that in combination with soil information, his landscapes seem to provide a framework for interpretation. His landscapes consisted of a high landform and a low landform connected by a scarp or steeper slope.

The first of these generalized landscapes consists of a smooth flat upland with a corresponding lowland. This model landscape compares well with the Atlantic Coastal Plain "red-edge" landscapes observed by Daniels and Gamble (1967). These soils in the southeastern states are well drained and hematitic often with a distinct reddish or yellowish colors. The wetter and more interior soils become progressively yellower first as a function of iron hydration and then gray due to iron losses from the poorly drained soils. We present a modified version here with soil classifications added to demonstrate the landscape–hydrology–soil continuum (Figure 3.39). The actual coastal area used for our model has a thin aquifer over an aquitard that is several miles wide. The hydraulic gradient is thus very low. The equipotential lines are widely spaced. Most of the recharge actually occurs from the Umbraquults to the Hapludults and not from the pocosin center muck-textured

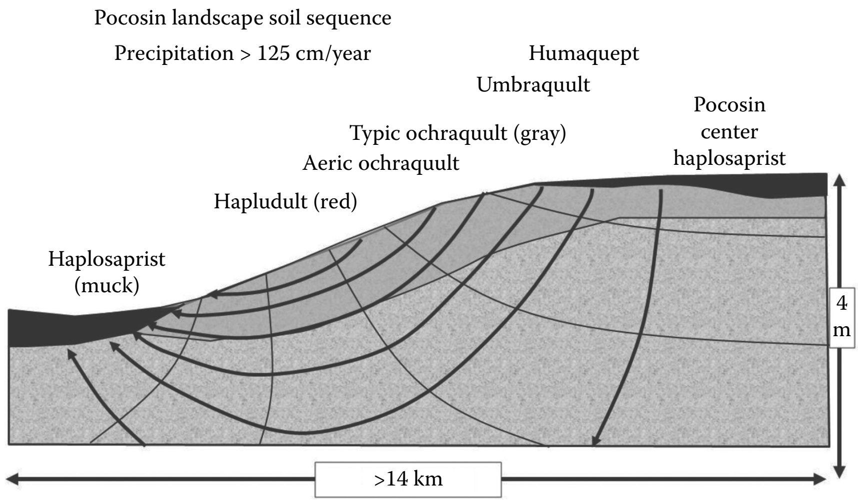


FIGURE 3.39
Soil distribution and flownet for a high rainfall flat upland typical of the low coastal plains near the Atlantic Ocean.

Histosol or organic soils. The pocosin center soils only receive rainwater as a water source (ombrotrophic) but drain the water exceedingly slowly such that the water becomes stagnant (stagno-groundwater recharge). The nutrients and soluble ions are slowly removed over time. The pocosin center soils, therefore, are mostly leached Histosols (organic soils). The Haplosaprist muck in the low landscape position in Figure 3.39 is an example of a mineralotrophic soil (mineral-rich Histosol).

Recharge is highest in the soils on the edge of the upper landform. These soils have argillic horizons and have lost iron due to reduction grading from the Hapludult to the Umbraquult. Colors range from red in the oxidized Hapludults to gray in the more reduced Umbraquults.

Winter's (1988) second generalized landscape, which he called "hummocky topography," is typified by local flow systems centered on depressions and intervening microhighs. We illustrate this type of landscape with a flownet modeled from an area in south central North Dakota (Figure 3.40). The landscape transect that we sampled has seven distinct

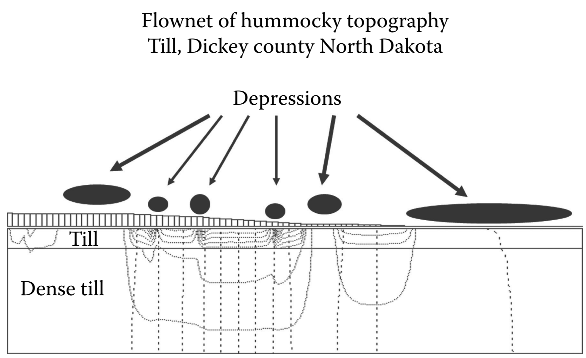


FIGURE 3.40
A FLOWNET simulation based on a landscape in till topography in south central North Dakota. The equipotential lines are 0.5 m decreasing increments from the high on the left (south) to the low on the right (north).

depressions with many smaller ones that are too small for the scale. The transect distance is about 3 km (2 miles) long. Equipotential lines occur in 0.5 m (20 in.) head intervals (dashed). There is approximately 6 m (20 ft) of head loss over the entire transect, with head decreasing from the left (south) to the right (north). Bold arrows mark the three largest wetlands. The illustration characterizes a landscape with regional flow being disrupted by complex local flow systems. At a larger scale, with the smaller depressions visible, flow is even more disrupted.

Lissey (1971) described depression-focused recharge and discharge ponds. Water in a ponded condition flows even if the movement is extremely slow. The movement impacts soils by removing or adding dissolved components and translocating clay materials. Discharging groundwater tends to add material to the soils, while recharging groundwater leaches material from the soil. Groundwater flow can reverse or alternate, thereby leading to a reversal in pedogenic processes. Over time, the dominant flow processes will be manifested in a unique pedogenic morphologic signature. An interpretation of the hydrologic regime can, therefore, be made using soil morphology (Richardson 1997).

A major problem with using soil morphology as an indicator of wetland hydrology, however, is that the natural groundwater hydrologic regime has often been altered through anthropogenic disturbance activities. These activities may include ditches and tile lines for removing water from a wetland, and dams and dikes that prevent water from entering a wetland. (Committee on Characterization of Wetlands 1995). It takes years for soil morphology to equilibrate with the new hydrologic regime. The morphologic indicators may be relict features indicative of the predisturbance hydrologic conditions.

For the examination of the small depressions that were too small to see individually in Figure 3.40, the smooth topography model of Winter (1988) could be utilized on each one because only local flow would be involved. For example in recharge wetlands, water collects in depressions and percolates slowly to the water table (Figure 3.41). Percolating water often forms mounded water tables in topographically low areas (Knuteson et al. 1989). Knuteson et al. (1989) described recharge wetlands formed in a subhumid climate of eastern North Dakota. They observed that the water table mounded under the depression during ponding events. The water table surface also had a steeper relief than existed on the

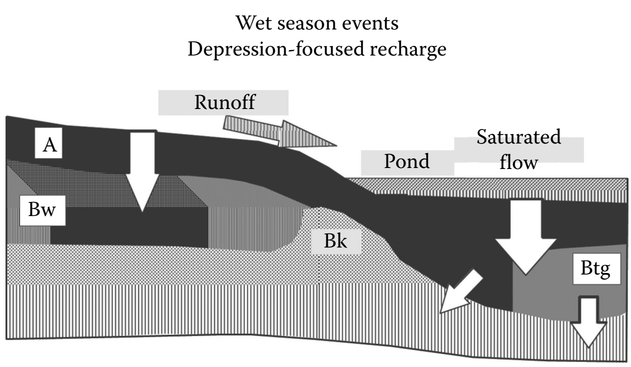


FIGURE 3.41
Wet season water flow system in depression-focused recharge wetlands. Variations in climate, stratigraphy, and topography alter details of the basic model. (After Lissey, A. 1971. *Geol. Assoc. Can. Spec. Pap.* 9: 333–341; Knuteson, J. A. et al. 1989. *Soil Sci. Soc. Am. J.* 53: 495–499; Richardson, J. L., J. L. Arndt, and J. Freeland. 1994. *Adv. Agron.* 52: 121–171.)

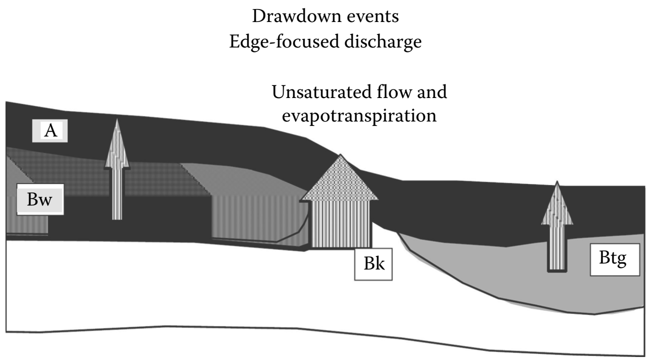


FIGURE 3.42
When the pond dries, upward flow is established by the drying influence at the surface of evapotranspiration and creates an upward wet to dry matric potential that initiates unsaturated upward flow. The edges of the depression have the longest period of time with upward flow and lack much downward flow in the wet periods, hence the thicker Bk horizons.

ground surface; the mound disappeared or was lowered during the drying of the wetland. Recharge wetlands are common in subhumid and drier climates, and they usually dry out during the growing season.

During precipitation events, or during spring snow melt, water moves by overland flow or by infiltration and throughflow into the wetland. The soil profiles tend to be leached in the uplands during these events, removing some carbonates and creating a Bw horizon. The Bw horizon is a weakly developed horizon (see Chapter 1).

The edge of the depression receives water that discharges from throughflow or transient flow during the aforementioned precipitation events (Figure 3.42). In times of low precipitation, these areas dry out and have abundant water moving upward via unsaturated flow through the soil in response to plant uptake and evapotranspiration. Dissolved materials are left as the water evaporates, resulting in the formation of Bk horizons. Carbonate levels in these horizons have been well in excess of 30%. This illustrates the fact that over one-quarter of the soil mass of these horizons has formed as an evaporite. Knuteson et al. (1989) examined the rate of formation of these horizons based on unsaturated flow and concluded that a horizon of this type can form in a few thousand years.

The pond area receives much water and temporarily has water above the soil surface nearly every year. The pond centers become inundated earlier and stay wet longer than other portions of the local landscape. Water moves downward through the profile along a hydraulic gradient (Figure 3.41), leaching and translocating material with it. Much of the dissolved material is completely leached from the profile, although some may be returned to the soil as the pond dries. Translocated clays accumulate at depth in the profile forming impermeable Btg horizons. These Btg horizons slow the percolation of water through the wetland bottom and increase the effectiveness of the pond to hold water.

The water flow system illustrated in Figures 3.41 and 3.42 results in soils with Bk horizons (carbonate accumulation) adjacent to soils with Bt horizons (carbonates removed and clays translocated). These soil types are extremely contrasting even though they are separated by only a few centimeters of elevation.

Zonation in Wetlands: Edge Effects

The edges of ponds and wetlands often display different flow regimes (e.g., saturated–unsaturated) that alternate several times per year. Such edge-focused processes were discussed in a preceding section. The wave-action mentioned earlier, for instance, created different landforms and soil types at the wetland edge. We previously mentioned the "red edge" effect and other edge phenomena. We will examine other edge-focused processes further in this section.

Flow reversals are specific hydrologic occurrences that are frequently observed at pond edges (Rosenberry and Winter 1997). Flow reversals occur when recharge flow changes to discharge flow, or vice versa. After rainfall events, infiltration and interflow shunt water to the pond edge and create a mounded water table (Figure 3.43). The water table is already near the soil surface at a pond edge. Groundwater moving as interflow now fills the pores that are not saturated. It is easy to saturate soils when the water table is near the surface both because of the thinness of the unsaturated zone and the large amount of unsaturated pore space present in the unsaturated capillary fringe (Winter 1983). The mounded water table at the wetland edge rises above the pond and acts as somewhat of a miniature drainage divide. The mound is a recharge mound, with groundwater moving both downslope into the pond and into the earth. The mound (Figure 3.43) intercepts interflow and shunts much of it via infiltration into the ground. Some of the interflow also recharges the mound. During these events, the soil is leached. This scenario is the opposite of the evaporative discharge often seen during dry periods at the edge, and the usual discharge of groundwater into the pond (Rosenberry and Winter 1997; Figure 3.44).

Plants at the edge of the wetland, such as phreatophytes and hydrophytes, are consumptive water users. Phreatophytes act like large water pumps, and selective plantings of these water users can alter local subsurface hydrology in the same manner as the pumping well in Figure 3.10. They create a depression in the water table, which illustrates that the water table mound is removed by water losses and replaced by a depression in the water table not long after the cessation of rain (Rosenberry and Winter 1997). The flow is reversed, and the water table depression also acts as a barrier to groundwater flowing into the pond. Wetland edges have frequent flow reversals of this type. During mound and depression phases, groundwater is restricted in its movement to the wetland.

Whittig and Janitzky (1963) in their classic paper described a wetland edge effect consisting of the accumulation of sodium carbonate (Figure 3.45). This type of edge effect has been widely known and is used as a model to illustrate salinization and alkalinization in warm

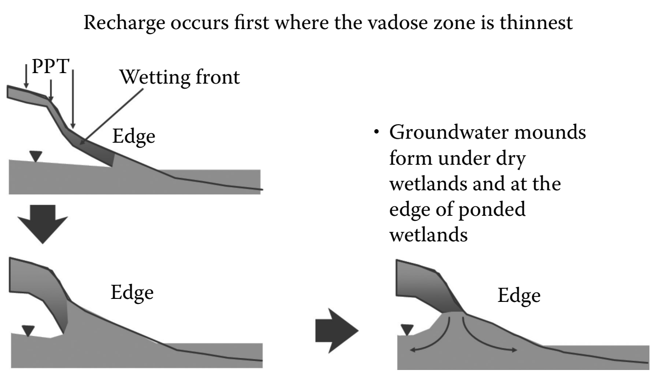


FIGURE 3.43
The development of a groundwater mound during rain events alters water flow into a wetland. The vadose zone is thinnest here. (After Winter, T. C. 1989. Hydrologic studies of wetlands in the northern prairies. In A. Van der Valk (Ed.) *Northern Prairie Wetlands*. Iowa State University Press, Ames, IA, pp. 16–54.)

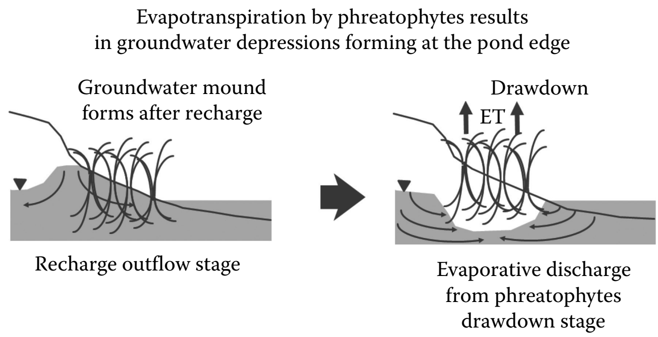


FIGURE 3.44
The mound dissipates quickly because the vegetations at wetland edges, particularly phreatophytes and hydrophytes, consume large quantities of water. These plants create a drawdown of the water table and disrupt water flow to the pond. Mounds alternating with drawdown depressions at the pond edge represent flow reversals.

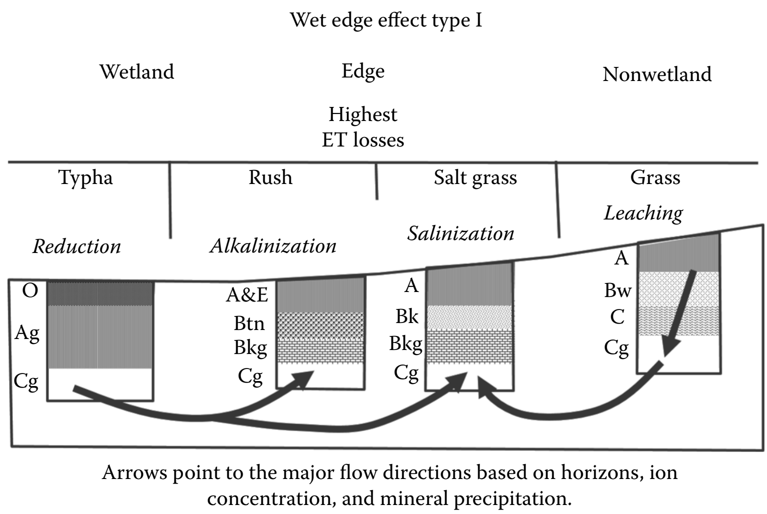


FIGURE 3.45
Edge-focused evaporative discharge with sodium carbonate development. This edge is more common in mesic and warmer climates. (After Whittig, L. D. and P. Janitzky. 1963. *J. Soil Sci.* 14: 322–333.)

climates. Chemical reduction via microbial transformations liberates the carbonate anion that then reacts with calcium to form the mineral calcite. Calcite precipitation removes calcium from the system, which increases the relative amounts of carbonate and bicarbonate anions in the soil solution. As the soil dries, matric potentials increase and water moves via capillarity transporting these anions, as well as sodium cations, toward the soil surface. During the evaporation process, the water loses dissolved carbon dioxide, resulting in an increase in pH carbonate forms when bicarbonate loses carbon dioxide. Whittig and Janitzky (1963) noted pH values as high as 10 in some of their profiles, with abundant sodium carbonate forming as a surface efflorescence. Inland and at slightly higher elevations, carbon dioxide is not a factor in carbonate formation. The carbon dioxide stays in solution, sulfate is not reduced, and thereby does not precipitate or form either calcium

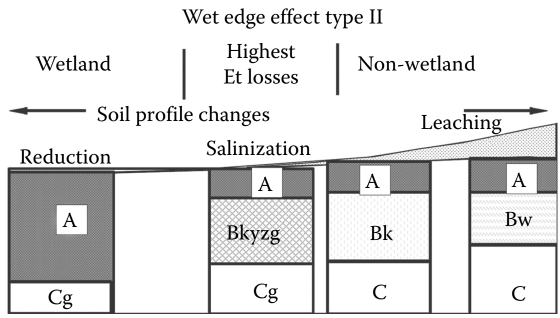


FIGURE 3.46
Evaporative discharge edge with gypsum and calcite rather than sodium carbonate. This edge is more common in cooler climates. (After Steinwand, A. L. and J. L. Richardson. 1989. *Soil Sci. Soc. Am. J.* 53: 836–842.)

carbonate or sodium carbonate. In these places, the soils become saline with accumulations of sodium and magnesium sulfates.

In northern climates, carbon dioxide remains in solution longer because the cool temperature retards sulfate reduction and allows for more dissolved carbon dioxide. In North Dakota and the Prairie Provinces of Canada, abundant sulfate is present and some reduction to sulfide occurs; however, the amount of carbonate in solution is less than the amount of available calcium (Arndt and Richardson 1988, 1989; Steinwand and Richardson 1989). Calcite and gypsum, therefore, are produced in place of sodium carbonate at the edge (Figure 3.46). The pathways of calcite and gypsum production are explained more fully in Chapter 15 and in Arndt and Richardson (1992). The result is that in northern areas, soil salinity is dominated by calcite and has pH levels that seldom exceed 8.3.

Wetland Hydrology and Jurisdictional Wetland Determinations

Regionalized Wetland Delineation Supplements

Wetlands are regulated under a variety of federal, and sometimes state, and local statutes. However, in order to regulate a resource, the resource must be defined. Prior to about 2005, the majority of the regulatory agencies that have jurisdiction over the nation's wetland resource used the 1987 U.S. Army Corps of Engineers (COE) manual to identify wetlands. The 1987 Wetland Delineation Manual, while discussing wetland hydrology in general terms, did not provide criteria that were specific enough to use in hydrological studies, and the Manual referenced poorly defined criteria such as "saturation to the surface for 5 to 10 percent of the growing season."

Current guidance involving hydrology criteria in the regionalized supplements to the 1987 Manual is more specific and incorporates advances occurring in wetland science over the past three decades. In particular, the commonly accepted saturation depth and duration criteria have been refined and specifically include saturation within the rooting zone as opposed to saturation to the soil surface with saturation including the capillary fringe. Linking saturation to the rooting zone acknowledges the importance of saturated conditions for maintaining hydrophytic vegetation and for developing hydric soils. A continuous saturation duration criterion for 14 days is the general standard for all regions, unless field data suggest otherwise. These standards, recommended by the Committee on Wetland Characterization (National Research Council 1995), have been generally incorporated into all regionalized supplements to the 1987 Manual.

While the current chapter provides the background and context to understand wetland hydrology and assessment, the supplements to the 1987 manual are the authorities that provide methods to assess the presence/absence of wetland hydrology in jurisdictional wetlands. Ten regional supplements exist, including:

- Alaska
- Arid West
- Atlantic
- Caribbean Islands Region
- Eastern Mountains and Piedmont
- Great Plains
- Hawaii and Pacific Islands
- Mid-west
- North-central and Northeast
- Western Mountains

The following discussion places the concepts of wetland hydrology into a regulatory context as described in the Supplements. The setting and condition of the wetland is particularly important when using the Supplements to evaluate wetland hydrology. Specifically:

- "Normal Circumstances" refer to the presence/absence of vegetation in the wetland. A wetland that has been cleared of vegetation is still a wetland if hydrophytic plants would return should the area be left to revegetate and return to "Normal Circumstances."
- "Problem Area" wetlands are naturally occurring wetland types that lack indicators of hydrophytic vegetation, hydric soil, or wetland hydrology, either periodically due to normal seasonal or annual variability, or permanently due to the nature of the soils or plant species on the site.
- "Atypical situations" are wetlands in which vegetation, soil, or hydrology indicators are absent due to recent human activities or natural events.

Wetland Hydrology Redefined

The U.S. Army COE (1987) Wetlands Delineation Manual defined wetland hydrology as follows:

> The term 'wetland hydrology' encompasses all hydrologic characteristics of areas that are periodically inundated or have soils saturated to the surface at some time during the growing season. Areas with evident characteristics of wetland hydrology are those where the presence of water has an overriding influence on characteristics of vegetation and soils due to anaerobic and reducing conditions, respectively. Such characteristics

> are usually present in areas that are inundated or have soils that are saturated to the surface for sufficient duration to develop hydric soils and support vegetation typically adapted for life in periodically anaerobic soil conditions
>
> **(paragraph 46, emphasis added)**

This definition did not provide sufficient information to define criteria that would be suitable to characterize wetland hydrology in areas where saturation or inundation are naturally absent for extended periods, or have been altered by drainage or other, generally man-induced mechanisms, respectively (see Section "Determining Wetland Hydrology when Delineation Conditions Do Not Reflect "Normal Circumstances," or Are in Atypical and Problem Area Wetlands"). The COE wetland delineation supplements now rely on regionalized hydrology indicators that are consistent with the general definition of wetlands.

However, a new technical standard has been developed (U.S. Army COE 2005), which is consistent with the recommendation of the National Committee on Wetland Characterization's recommendations (National Resource Council 1995).

> The site is inundated (flooded or ponded) or the water table is ≤12 inches below the soil surface for ≥14 consecutive days during the growing season at a minimum frequency of 5 years in 10 (≥50% probability). Any combination of inundation or shallow water table is acceptable in meeting the 14-day minimum requirement. Short-term monitoring data may be used to address the frequency requirement if the normality of rainfall occurring prior to and during the monitoring period each year is considered.
>
> **(U.S. Army COE, 2005, p. 12)**

The guidance provided in the Supplements relies on a larger suite of regionally appropriate hydrology indicators for use in "normal" wetland delineation situations that are not Atypical or Problem area wetlands and where normal circumstances exist. The supplement guidance is clear in indicating that field hydrologic indicators take precedence under most normal wetland delineations. The technical standard provided above defines wetland hydrology for use in atypical and/or problem area delineations or where normal circumstances do not exist.

Depth and Duration of Inundation or Saturation

Under the Technical Standard, saturation to the surface for some period is no longer a requirement for wetland hydrology to be present. It must be clear that the current Technical Standard has not changed the definition of wetland hydrology, nor does it result in an overall change in the wetland boundary. However, it is more consistent with the general definition of wetlands, and is based on additional field experience and consensus of experts.

References to saturation to the surface have been replaced by the requirement for the presence of a water table at or less than 12-in. (30 cm) of the surface (the major portion of the rooting zone) for a period of 14 consecutive days during normal years (those years with normal precipitation for 5 years out of 10%, or 50% of the time) unless different depth-duration standards have been adopted at a local or regional level. All of the 10 regional supplements as of the date of this writing use the presence of a water table at or above 12 in. for 14 consecutive days in most years as the generally applicable standard for use in problem and atypical area wetland delineations.

The change in depth requirement from to a water table at or above 12 in. acknowledges the presence of a capillary fringe, but does not emphasize it, nor does it require a difficult and potentially arbitrary determination of the effect of the capillary fringe on the presence/absence of wetland hydrology. Furthermore, the technical standard does not differentiate between sandy and loamy soils which referred to the assumed greater thickness of the capillary fringe in finer when compared to coarser textured soils. Inclusion of the capillary fringe based on soil texture had numerous qualifiers that compromised the use of the capillary fringe in soils with differing textures (see Section "Adhesion, Cohesion, and Capillarity" for additional information). The updated Technical Standard for wetland hydrology provides more utility in defining the wetland hydrologic criterion in situations where data may be directly used because normal indicators are missing in problem and atypical areas, and is more consistent with field observations.

In addition, saturation duration criteria have been increased from the poorly defined standards provided in Table 5 of the 1987 Manual which indicated that areas that transition from wetland to upland are irregularly inundated or saturated for greater than 5%–12.5% of the time during the growing season, and qualified this range by stating "many areas having these characteristics are not wetlands." In practice, areas that were saturated to the surface for >5% of the growing season were considered to meet the wetland hydrology criterion. For example, in the Minneapolis, Minnesota area, the growing season lasts from about May 1 to October 1, or 153 days based on the soil survey data from the area; 5% of the growing season equates to 7.65 days. Thus, in Minneapolis, inundation or saturation to the surface must be present for an absolute minimum of 8 days during the growing season for wetland hydrology to exist as defined in the 1987 manual.

However, the new Technical Standard requires a water table be continuously present at or above 12 in. depth for 14 days, regardless of the length of the growing season unless regional data indicates a shorter (or longer) duration in most years. Under normal circumstances and where the wetland is not a problem or atypical wetland, field indicators of wetland hydrology, wetland vegetation, and wetland soils are preferred over a field study of wetland hydrology.

Field Methodology for Determining Wetland Hydrology: Emphasis on Indicators

Field methodology to determine the presence/absence of wetland hydrology has changed considerably from the methods proposed in the 1987 Manual. While it has always been recognized that hydrology is the most transient, dynamic, and difficult to determine of the three wetland indicators, the guidance provided in the supplements:

- Regionalizes field indicators of wetland hydrology, and generally provides more indicators when compared to the indicators in the 1987 Manual.
- Places an emphasis on primary and secondary field indicators in situations where normal circumstances prevail, and the setting is not an atypical or problem area setting.
- Specifically identifies atypical and problem areas that will require additional methods to describe and define the presence of wetland hydrology.
- Describes methods that have been specifically developed and tested for use in evaluating the presence/absence of wetland hydrology by regions and sub-regions covered by a specific supplement.

Moreover, in most cases the number of field indicators of wetland hydrology provided in the regionalized supplements has doubled when compared to those provided in the

1987 Manual, the supplement indicators are regionally specific, and one primary and two secondary indicators are required.

Direct Determination of Water Table Depth

The U.S. Army COE 1987 manual provided a field methodology for determining if soil saturation is present:

> Examination of this indicator requires digging a soil pit to a depth of 16 inches and observing the level at which water stands in the hole after sufficient length of time has been allowed for water to drain into the hole. The required time will vary depending on soil texture. In some cases, the upper level at which water is flowing into the pit can be observed by examining the wall of the hole. This level represents the depth to the water table. The depth to saturated soils will always be nearer the surface due to the capillary fringe. For soil saturation to impact vegetation, it must occur within a major portion of the root zone (usually within 12 inches of the surface) of the prevalent vegetation.
>
> **(paragraph 49.b. [2])**

This open borehole methodology indicates that the parameter being measured is whether the water table is within 12 in. of the surface and is consistent with the guidance under the Technical Standard, and both the 1987 Manual and regionalized supplements.

Primary and Secondary Indicators of Hydrology

Regionalization of wetland delineation techniques has resulted in a greater number of regionally specific indicators of wetland hydrology, increased attention to hydrogeomorphic setting as a primary characteristic of wetlands and determination of "Normal Circumstances," and identification of "Atypical" and "Problem Area" wetland settings. In the supplements, wetland hydrology indicators are presented in four groups.

Indicators in Group A (Table 3.2) are based on the direct observation of surface water or groundwater during a site visit. Group B consists of evidence that the site is subject to flooding or ponding, although it may not be inundated currently. These indicators include water marks, drift deposits, sediment deposits, and similar features. Group C consists of other evidence that the soil is saturated currently or was saturated recently. Some of these indicators, such as oxidized rhizospheres surrounding living roots and the presence of reduced iron or sulfur in the soil profile, indicate that the soil has been saturated for an extended period. Group D consists of landscape and vegetation characteristics that indicate contemporary rather than historical wet conditions. Wetland hydrology indicators are intended as one-time observations of site conditions that are sufficient evidence of wetland hydrology.

Within each group, indicators are divided into two categories—primary and secondary—based on their estimated reliability in this region. Primary indicators provide stand-alone evidence of a current or recent hydrologic event; some of these also indicate that inundation or saturation was long-lasting. Secondary indicators provide evidence of recent inundation or saturation when supported by one or more other primary or secondary wetland hydrology indicators, but should not be used alone. One primary indicator from any group is sufficient to conclude that wetland hydrology is present; the area is a wetland if indicators of hydric soil and hydrophytic vegetation are also present. In the absence of a primary indicator, two or more secondary indicators from any group are required to conclude that wetland hydrology is present.

TABLE 3.2

Wetland Hydrology Indicators for the Midwest Region[a]

	Category	
Indicator	**Primary**	**Secondary**
Group A—Observation of Surface Water or Saturated Soils		
A1—Surface water	X	
A2—High water table	X	
A3—Saturation	X	
Group B—Evidence of Recent Inundation		
B1—Water marks	X	
B2—Sediment deposits	X	
B3—Drift deposits	X	
B4—Algal mat or crust	X	
B5—Iron deposits	X	
B7—Inundation visible on aerial imagery	X	
B8—Sparsely vegetated concave surface	X	
B9—Water-stained leaves	X	
B13—Aquatic fauna	X	
B14—True aquatic plants	X	
B6—Surface soil cracks		X
B10—Drainage patterns		X
Group C—Evidence of Current or Recent Soil Saturation		
C1—Hydrogen sulfide odor	X	
C3—Oxidized rhizospheres along living roots	X	
C4—Presence of reduced iron	X	
C6—Recent iron reduction in tilled soils	X	
C7—Thin muck surface	X	
C2—Dry-season water table		X
C8—Crayfish burrows		X
C9—Saturation visible on aerial imagery		X
Group D—Evidence from Other Site Conditions or Data		
D9—Gauge or well data	X	
D1—Stunted or stressed plants		X
D2—Geomorphic position		X
D5—FAC-neutral test		X

[a] U.S. Army COE 2010.

Wetland indicators for the Midwest Region are provided in Table 3.2 however, the reader is directed to the applicable regional supplement for additional information on hydrology indicators that are specific for the region of interest.

Determining Wetland Hydrology When Delineation Conditions Do Not Reflect "Normal Circumstances," or Are in Atypical and Problem Area Wetlands

The 1987 Manual made it clear that the presence of wetland hydrology may not be inferred from the presence of hydric soils and a predominance of hydrophytic plants, particularly when an area has been altered from "normal circumstances." The 1987 manual states that:

> … sole reliance on vegetation or either of the other parameters as the determinant of wetlands can sometimes be misleading. Many plant species can grow successfully in both wetlands and non-wetlands, and hydrophytic vegetation and hydric soils may persist for decades following alteration of hydrology that will render an area a non-wetland.
>
> **(paragraph 19)**

Moreover, the 1987 Manual identified methodology that would be applicable to Problem Area and Atypical wetlands (Section F, the 1987 Manual).

The Supplements consistently address difficult wetland delineation conditions in a separate, dedicated chapter (typically Chapter 5) titled "Difficult Wetland Situations…." For example, the Midwest region is glaciated and dominated by agricultural land use with wetlands typically existing as a mosaic of undrained depressions of varying degrees of permanence emplaced at varying elevations in glacial till. The Midwest Supplement (U.S. Army COE 2010) discusses the most common difficult wetland delineation settings as

- Agricultural lands (Atypical Settings)
- Problematic Hydrophytic Vegetation (Problem Area Wetlands)
- Problematic Hydric Soils (Problem Area Wetlands)
- Wetlands that Periodically Lack Indicators of Wetland Hydrology (both Atypical and Problem Area Wetlands
- Wetland/non-wetland mosaics (Problem Area Wetlands)

With respect to wetland hydrology, methods are provided that can be used to characterize wetland hydrology in these situations, where strict reliance on hydrology indicators may give erroneous delineation results.

Many of these techniques use the approach and information provided above in Sections "Review of Basic Hydrologic Principles," "Soils, Water, and Wetlands," and "Applications: Wetland Hydrology," above. The effects of the type and magnitude of the alterations and their relationship to preexisting conditions are supported by the methods summarized below and in this chapter. The various methods employed to refine the hydrological assessment of a site should converge on a wetland determination that is supported by data and the best professional judgment of the delineator. It is important when considering wetland hydrology in difficult situations that the delineation narrative includes the following in dataform notes or in the delineation text.

Describe the type of alteration. Anthropogenic impacts to wetland hydrology may be subtle or obvious, and may result in creation of wetter or drier conditions. Agricultural drainage ditches, drain tiles, dikes, levees, and filling are obvious attempts to remove water from an area or prevent water from flowing onto an area. Stormwater drains and diversions are obvious indicators that water may be added to an area. Other effects of urbanization and agricultural use, including effects of off-site activities, are more subtle, and may have broad, regional impacts on the groundwater system that are not obvious, yet may result in a continuous, overall decline in the health and magnitude of the wetland resource.

Describe the effects of the alteration. The effects of several hydrologic alterations can be theoretically addressed by employing many of the concepts examined in this chapter, focusing on an assessment of the effects the alterations have on the water balance of the study area.

Characterize the preexisting conditions. This characterization is commonly performed with an interpretation of the existing aerial photo history augmented with map analyses, literature searches, soil survey information, and soils and vegetation documentation. An important change that should be mentioned is the change from phreatophytes, which are heavy water users, to field crops, which use very little water comparatively.

The following examples come from the Midwest Supplement (U.S. Army COE 2010).

Atypical Conditions: Agricultural Lands in the Midwest Region

"Farmed Wetlands" represent Atypical settings where hydrological characteristics of wetlands may be missing or altered due to the action of man, either through diversion, land leveling, and/or subsurface and surface drainage or irrigation. They may also represent "Problem Areas," and situations where conditions are "Not normal" circumstances. To determine the presence/absence of wetland hydrology in agricultural lands, the Midwest Supplement provides the following guidance. The reader is directed to the Midwest Supplement (U.S. Army COE 2010) for greater detail on specific methods.

1. Examine the site for existing indicators of wetland hydrology.
2. Examine five or more years of annual Farm Service Agency aerial photographs, or aerial photos from other sources, for wetness signatures listed in Part 513.30 of the National Food Security Act Manual (USDA NRCS 1994) or in wetland mapping conventions available from NRCS offices or online in the electronic Field Office Technical Guide (eFOTG) (http://www.nrcs.-usda.gov/technical/efotg/).
3. Estimate the effects of ditches and subsurface drainage systems using scope-and-effect equations (USDA NRCS 1997). A web application to analyze data using various models is available at http://www.wli.nrcs.usda.gov/technical/web_tool/-tools_java.html
4. Use state drainage guides to estimate the effectiveness of an existing drainage system (USDA NRCS 1997).
5. Use hydrologic models (e.g., runoff, surface water, and groundwater models) to determine whether wetland hydrology is present (USDA NRCS 1997).
6. Monitor the hydrology of the site in relation to the appropriate wetland hydrology technical standard (U.S. Army COE 2005).

The presence of existing indicators of wetland hydrology may be relict in areas with altered hydrological conditions.

Wetlands That Periodically Lack Indicators of Wetland Hydrology

If the site is visited during a time of normal precipitation amounts and it is inundated or the water table is near the surface, then the wetland hydrology determination is straightforward. During the dry season, however, surface water may recede from wetland margins, water tables may fall below levels characteristic of wetlands, or wetlands may dry out completely. Superimposed on this seasonal cycle is a long-term pattern of multi-year droughts alternating with years of higher-than-average rainfall. For example, some wetlands in the

Midwest do not become inundated or saturated in some years and, during drought cycles, may not inundate or saturate for several years in a row, in spite of the fact that the technical standard for the presence of wetland hydrology would be met when considered over a long term (e.g., >50 of 100 years).

The evaluation of presence/absence of wetland hydrology should consider the timing of the site visit in relation to normal seasonal and annual hydrologic variability, and whether the amount of rainfall prior to the site visit is typical. The Midwest Supplement describes a number of approaches that can be used to determine whether wetland hydrology is present on sites where indicators of hydrophytic vegetation and hydric soil are present but hydrology indicators may be lacking due to normal variations in rainfall or runoff, human activities that destroy hydrology indicators, and other factors.

To determine the presence/absence of wetland hydrology when hydrophytic vegetation and hydric soil indicators are present, the Midwest Supplement provides the following guidance. The reader is directed to the Midwest Supplement (U.S. Army COE 2010) for greater detail on specific methods.

1. Verify that indicators of hydrophytic vegetation and hydric soil are present or absent due to disturbance or other problem situations.

   Data forms in all Supplements require a determination of whether or not "Normal Circumstances" prevail, and an on-site evaluation of whether or not the potential wetland area being considered is Atypical or a Problem Area. If the answer is yes to any, the applicable sections in the chapter on "Difficult Wetlands" should be consulted.

2. Verify that the site is in a geomorphic position that is likely to collect or concentrate water.

   a. Concave surface (e.g., depression or swale)
   b. Active floodplain or low terrace
   c. Level or nearly level area (e.g., 0%–3% slope)
   d. Toe slope (Figure 3.13) or an area of convergent slopes (Figures 3.14 and 3.15)
   e. Fringe of another wetland or water body
   f. Area with a restrictive soil layer or aquitard within 24 in. (60 cm) of the surface
   g. Area where groundwater discharges (e.g., a seep)
   h. Other (explain in field notes why this area is likely to be inundated or saturated for long periods)

   Landform, local relief (slope morphology), and slope percent are now components of the site data form that need to be completed and evaluated in the assessment of wetland hydrology. Moreover, geomorphic position (characteristic of regional wetland settings) is frequently a secondary indicator of hydrology.

   Geomorphic positions such as landform (e.g., hillslope, terrace, undrained depression, floodplain), local relief (concave, convex, and linear slopes), and slope percent are now components that must be provided on delineation dataforms that are provided for each regional Supplement. In the Midwest Supplement (U.S. Army COE 2010), geomorphic position is a secondary indicator of hydrology if the position is in an area that concentrates water, for example, depression, drainageway, concave position on a floodplain, the toe of slope, low elevation fringe of a pond or other waterbody, or an area where groundwater discharges.

3. Use one or more of the following approaches to determine whether wetland hydrology is present and the site is a wetland.
    a. Site visits during the dry season
    b. Periods with below-normal rainfall
    c. Drought years
    d. Reference sites
    e. Hydrology tools
    f. Evaluate multiple years of aerial photography
    g. Employ long-term hydrologic monitoring)

Considerations, Caveats

Jurisdictional wetland delineation has as its focus the dry edge of the wetland. It is an unfortunate reality that wetland delineation does not focus on wetland presence or absence, but instead focuses on the areal extent of the wetland. The term "unfortunate" is used because wetland delineation takes the most dynamic portion of the wetland that exists as a transition zone and turns it into a two-dimensional line. It is for these reasons that most of the disputes involving jurisdictional wetland boundaries occur at the wetland edge: we take something that exists as a gradient in three dimensions and turn it into two. In many situations, this representation of the wetland boundary is unrealistic.

It is also at this dry edge where the soil–landscape–hydrology interactions result in the development of hydric soil morphology that is transitional to upland soil characteristics. In addition to being the location of the jurisdictional boundary, sediment deposition also occurs primarily at the wetland edge. Sediment deposition has significant impacts on wetland longevity, functions, and quality, especially when accelerated by human activities. It is unfortunate that researchers often ignore these transitional areas. Pond interiors are often the only locations that have water level recorders and other instrumentation for measuring hydroperiod. Measuring hydroperiod only in the interiors and not on the wetland edges results in an incomplete picture of hydroperiod. It is only through an understanding of the dynamic hydrology of the transition zone between wetland and upland that we can understand the interactions between hydrology, soils, and vegetation sufficiently to make accurate jurisdictional determinations, and wisely manage the wetland resource.

In recent years, the extent of wetland jurisdiction under the Clean Water Act has been challenged in the federal court system, leading to multiple U.S. Supreme Court decisions[*] that have implications for certain wetlands. In particular, the U.S. Supreme Court has ruled that wetlands may, in some cases, be subject to regulation under the Clean Water Act even if they are not immediately bordering or neighboring other jurisdictional waters (i.e., non-adjacent wetlands). The wetlands in question must be capable of exerting hydrological or ecological influence on jurisdictional waters. The Corps of Engineers and EPA have established criteria for determining whether a wetland exerts such influence,[†] and these criteria acknowledge that hydrological processes, both above ground and

[*] Solid Waste Agency of Northern Cook County v. United States 2001; *Rapanos v. United States* 2006; *Carabell v. United States* 2006.

[†] US Army COE and US EPA, Memorandum: Clean Water Act Jurisdiction Following the US Supreme Court Decision in *Rapanos v. United States* and *Carabell v. United States*, June 6, 2007.

belowground, may play an important role in connecting wetlands to other bodies of water. Given the often transient nature of hydrological flows, however, the current criteria are not always easy to apply. As a result, practitioners and regulators have sometimes faced difficulty and ambiguity in demonstrating what Justice Kennedy termed a "significant nexus" between non-adjacent wetlands and other jurisdictional waters. The Corps of Engineers and EPA continue to seek scientific input on the hydrological nature and ecological implications of connectivity between wetlands and downstream waters in order to refine and clarify regulations pertaining to the jurisdiction of wetlands under the Clean Water Act.

Summary

A wetland, as suggested by the nature of the name, consists of two natural media interacting: water and soil. Wetland hydrology is dynamic and can change with a single rainstorm event, or a rapid snowmelt, or during a hot windy day. The wetland water balance is the fundamental relationship between inputs, outputs, and storage that dictates the presence or absence of a wetland. The water may come from the landscape where it has been gathered from its catchment basin or fall directly on the wetland via precipitation. Water, once in the wetland, either stays, leaves by evapotranspiration, or it drains away either vertically or laterally.

To be a hydric soil, the soil must remain saturated for an extended time and be chemically reduced. The chemical and physical processes that occur by water moving into, through, and from the soil alter it in distinct, visible ways. These changes occur slowly over time as a response to the water activity. This visible hydrologic signature is called soil morphology. Recharge dominance, for instance, is the direct movement of water from the wetland to groundwater. The movement of water over time in this manner leaches soluble material and translocates clay in the soil. Discharge dominance, on the other hand, adds materials such as calcium carbonate to hydric soils. Iron is usually chemically reduced in saturated conditions and often alternatively oxidized during drier periods. This creates a distinct morphological pattern that reflects both the soil chemistry and hydrologic conditions. Hydric soil indicators developed from the process.

Landscape, climatological, and biological conditions must exist to get and keep a wetland wet. Hillslope geometry and position, such as the base of long slopes, shed and concentrate water at certain places. Depressions frequently constrain water from flowing freely to a stream. Strata, such as sand lenses, may gather the water from a large catchment and concentrate the water in a wetland. Climatic constraints, such as copious quantities of precipitation or very low evapotranspiration rates, maintain water in the wetland throughout a year or periodically during a wet season. Certain plants may foster the retention of water and aid in wetland creation. All these conditions are reflected in hydric soils. The soils reflect the hydrology of the pedons throughout the wetlands and can be used to determine the hydrology expected over time, the wetland as a whole, or zones within a wetland.

Alteration of the wetland, frequently for an economic purpose, changes wetland hydrology. Sadly, a rather long period of time may occur before the hydric soils equilibrate and reflect the new hydrologic conditions via their soil morphology.

References

Arndt, J. L. and J. L. Richardson. 1988. Hydrology, salinity, and hydric soil development in a North Dakota prairie-pothole wetland system. *Wetlands* 8: 94–108.

Arndt, J. L. and J. L. Richardson. 1989. Geochemical development of hydric soil salinity in a North Dakota prairie-pothole wetland system. *Soil Sci. Soc. Am. J.* 53: 848–855.

Arndt, J. L. and J. L. Richardson. 1992. Carbonate and gypsum chemistry in saturated, neutral pH soil environments. In R. D. Robarts, and M. L. Bothwell (Eds.) *Aquatic Ecosystems in Semi-arid Regions: Implications for Resource Management. N. H. R. I. Symposium Series*, Vol. 7, Environment Canada, Saskatoon, Saskatchewan, Canada, pp. 179–187.

Arndt, J. L. and J. L. Richardson. 1993. Temporal variation in salinity of shallow groundwater collected from periphery of North Dakota wetlands (U.S.A.). *J. Hydrol.* 141: 75–105.

Bailey, R. W. and O. L. Copeland. 1961. Low flow discharges and plant cover relations on two mountain watersheds in Utah. *Int. Assoc. Sci. Hydrol. Publ.* 51: 267–278.

Bates, B. C., Z. W. Kundzewicz, S. Wu, and J. P. Palutikof (Eds.) 2008. *Climate Change and Water.* Technical Paper of the Intergovernmental Panel on Climate Change, IPCC Secretariat, Geneva, 210 pp.

Beven, K. and P. Germann. 2013. Macropores and water flow in soils revisited. *Water Resour. Res.* 49: 3071–3092.

Beven, K. J. 1997. TOPMODEL: A critique. *Hydrol. Process.* 11: 1069–1085.

Beven, K. J. and M. J. Kirkby. 1979. A physically based, variable contributing area model of basin hydrology. *Hydrol. Sci. Bull.* 24(1): 43–69.

Bouma, J. 1990. Using morphometric expressions for macropores to improve soil physical analyses of field soils. *Geoderma* 46: 3–11.

Brady, N. L. and R. R. Weil. 1998. *The Nature and Properties of Soils*, 12th ed. Prentice-Hall, Englewood Cliffs, NJ, 740 pp.

Brinson, M. M. 1993. *A Hydrogeomorphic Classification for Wetlands*. Wetlands Research Program Tech. Rept. WRP-DE-4. U.S. Army Corps of Engineers, Waterways Experiment Station, Vicksburg, MS.

Brooks, R., M. Brinson, K. Havens, C. Hershner, R. Rheinhardt, D. Wardrop, D. Whigham, A. Jacobs, and J. Rubbo. 2011. Proposed hydrogeomorphic classification for wetlands of the Mid-Atlantic Region, USA. *Wetlands* 31: 207–219.

Carroll, D. 1970. *Rock Weathering*. Plenum Press, New York.

Cedargren, H. R. 1967. *Seepage, Drainage, and Flow Nets*. John Wiley & Sons, New York.

Chorley, R. J. 1978. The hillslope hydrological cycle. In M. J. Kirkby (Ed.) *Hillslope Hydrology*. John Wiley & Sons, New York, pp. 1–42.

Committee on Characterization of Wetlands. 1995. *Wetlands: Characteristics and Boundaries*. W.W. Lewis, (Chair), National Research Council, National Academy of Sciences, Washington, DC.

Cooper, D. J., J. S. Sanderson, D. I. Stannard, and D. P. Groeneveld. 2006. Effects of long-term water table drawdown on evapotranspiration and vegetation in an arid region phreatophyte community. *J. Hydrol.* 325: 21–34.

Cowardin, L. M., V. Carter, F. C. Golet, and E. T. LaRoe. 1979. *Classification of Wetland and Deepwater Habitats of the United States*. FWS/OBS-79/31. U.S. Fish and Wildlife Service, Washington, DC.

Daniels, R. B. and E. E. Gamble. 1967. The edge effect in some Ultisols in the North Carolina coastal plain. *Geoderma* 1: 117–124.

Department of Land Conservation and Development. 2009. *Climate Ready Communities: A Strategy for Adapting to Impacts of Climate Change on the Oregon Coast*. Coast Oregon Coastal Management Program Salem, Oregon.

Desmet P. J. J., J. Poesen, G. Govers, and K. Vandaele. 1999. Importance of slope gradient and contributing area for optimal prediction of the initiation and trajectory of ephemeral gullies. *Catena* 37: 377–392.

Elburg, H. van, G. B. Engelen, and C. J. Hemker. 1990. *Flownet Users Manual*. The Free University, Institute of Earth Sciences, DeBoeleaan 1085, Amsterdam, the Netherlands.

Freeland, J. A. 1996. *Soils and sediments as indicators of agricultural impacts on northern prairie wetlands*. PhD dissertation, North Dakota State University, Fargo, ND.
Freeland, J. A., J. L. Richardson, and L. A. Foss. 1999. Soil indicators of agricultural impacts on northern prairie wetlands: Cottonwood Lake Research area, North Dakota. *Wetlands* 19: 78–89.
Freeze, R. A. and J. A. Cherry. 1979. *Groundwater*. Prentice-Hall, Englewood Cliffs, NJ.
Freeze, R. A. and P. A. Witherspoon. 1966. Theoretical analysis of regional groundwater flow: 1. Analytical and numerical solutions to the mathematical model. *Water Resour. Res*. 2: 641–656.
Freeze, R. A. and P. A. Witherspoon. 1967. Theoretical analysis of regional groundwater flow: 2. Effect of water table configuration and subsurface permeability variation. *Water Resour. Res*. 3: 623–634.
Freeze, R. A. and P. A. Witherspoon. 1968. Theoretical analysis of regional groundwater flow: 3. Quantitative interpretations. *Water Resour. Res*. 4: 581–590.
Gambrel, R. P. and W. H. Patrick, Jr. 1978. Chemical and microbiological properties of anaerobic soils and sediments. In D. D. Hook, and R. M. M. Crawford (Eds.) *Plant Life in Anaerobic Environments*. Ann Arbor Sci. Publ. Inc., Ann Arbor, MI, pp. 375–423.
Gilley, J. E., B. D. Patton, P. E. Nyren, and J. R. Simanton. 1996. Grazing and haying effects on runoff and erosion from a former conservation reserve program site. *Appl. Eng. Agric.* 12: 681–684.
Gitay, H., C. Max Finlayson, and N. Davidson. 2011. *A Framework for Assessing the Vulnerability of Wetlands to Climate Change*. Ramsar Technical Report No. 5/CBD Technical Series No. 57. Ramsar Convention Secretariat, Gland, Switzerland & Secretariat of the Convention on Biological Diversity, Montreal, Canada. ISBN 92-9225-361-1 (print); 92-9225-362-X (web).
Greenwood, D. J. 1961. The effect of oxygen concentration on the decomposition of organic materials in soil. *Plant Soil* 14: 360–376.
Hartig, E. K., V. Gornitx, A. Kolker, F. Mushake, and D. Fallon. 2002. Anthropogenic and climate-change impacts on salt marshes of Jamaica Bay, New York City. *Wetlands* 22(1): 71–89.
Heath, R. C. 2004. *Basic Ground-Water Hydrology 10th Printing (Revised)*. United States Geological Survey Water Supply Paper 2220.
Hewlett, J. D. and W. L. Nutter. 1970. The varying source area of streamflow from upland basins. *Paper Presented at Symposium on Interdisciplinary Aspects of Watershed Management*, Montana State University, American Society of Civil Engineers, New York, pp. 65–83.
Hornberger, G. M., J. P. Raffensperger, P. L. Wilberg, and K. Eshleman. 1998. *Elements of Physical Hydrology*. John Wiley & Sons, New York.
Hubbert, M. K. 1940. The theory of groundwater motion. *J. Geol*. 48: 785–944.
Jenny, H. 1941. *Factors of Soil Formation*. McGraw-Hill, New York.
Johnson, W. C., B. V. Millett, T. Gilmanov, R. A. Voldseth, G. R. Guntenspergen, and D. E. Naugle. 2005. Vulnerability of northern prairie wetlands to climate change. *Bioscience* 55: 863–872.
Joosten, H., M.-L. Tapio-Biström, and S. Tol (Eds.) 2012. *Peatlands—Guidance for Climate Change Mitigation through Conservation, Rehabilitation and Sustainable Use*. Food and Agriculture Organization of the United Nations and Wetlands International.
Kirkby, M. J. and R. J. Chorley. 1967. Throughflow, overland flow and erosion. *Bull. Int. Assoc. Sci. Hydrol.* 12: 5–21.
Kirkham, D. 1947. Studies of hillslope seepage in the Iowan drift area. *Soil Sci. Soc. Am. Proc*. 12: 73–80.
Knuteson, J. A., J. L. Richardson, D. D. Patterson, and L. Prunty. 1989. Pedogenic carbonates in a Calciaquoll associated with a recharge wetland. *Soil Sci. Soc. Am. J.* 53: 495–499.
Kramer, J., J. Printz, J. Richardson, and G. Goven. 1992. Managing grass, small grains, and cattle. *Rangelands* 14(4): 214–215.
Kutilek, M. and D. R. Nielsen. 1994. *Soil Hydrology*. Catena Verlag, Cremlingen-Destedt, Germany.
Lehmann, P. and D. Or. 2009. Evaporation and capillary coupling across vertical textural contrasts in porous media. *Phys. Rev. E* 80: 046318.
Lissey, A. 1971. Depression-focused transient groundwater flow patterns in Manitoba. *Geol. Assoc. Can. Spec. Pap.* 9: 333–341.

Maasland, M. and Haskew, H. C. 1957. The auger hole method of measuring the hydraulic conductivity of soil and its application to tile drainage problems. In *Proceedings, Third Congress on International Commission on Irrigation and Drainage*, San Francisco, Part I-R5, Question 8, pp. 8.69–8.114.

Mausbach, M. J. 1992. Soil survey interpretations for wet soils. In J. M. Kimble (Ed.) *Proc. 8th Intern. Soil Correlation Meeting (VIII ISCOM): Characterization, Classification and Utilization of Soils*. USDA-SCS. National Soil Survey Center, Lincoln, NE.

Meyboom, P. 1967. Mass-transfer studies to determine the groundwater regime of permanent lakes in hummocky moraine of western Canada. *J. Hydrol*. 5: 117–142.

Miller, J. J., D. F. Acton, and R. J. St. Arnaud. 1985. The effect of groundwater on soil formation in a morainal landscape in Saskatchewan. *Can. J. Soil Sci*. 65: 293–307.

Mills, J. G. and M. Zwarich. 1986. Transient groundwater flow surrounding a recharge slough in a till plain. *Can. J. Soil Sci*. 66: 121–134.

Mohring, E. 2011. *Evaluating Antecedent Precipitation Conditions for Assessing Wetland Hydrology Using Climate Data Available in Minnesota*. Minnesota Boards of Water and Soil Resources Guidance. Minnesota Board of Soil and Water Resources, St. Paul, MN.

Moore, P. D. and D. J. Bellamy. 1974. *Peatlands*. Springer Verlag, New York.

Moorhead, K. K. and M. M. Brinson. 1995. Response of wetlands to rising sea level in the lower coastal plain of North Carolina. *Ecol. Appl*. 5: 261–271.

National Research Council. 1995. *Wetlands: Characteristics and Boundaries*. The National Academies Press, Washington D.C.

National Wetlands Working Group. 1988. *Wetlands of Canada*. *Ecological Land Classification Series*, Vol. 24. Environment Canada Polyscience Publications, Ottawa, Ontario.

Noble, C. V. 2006. *Water Table Monitoring Project Design*. ERDC TN-WRAP-06-2. U.S. Army COE Research and Development Center, Visckburg, MS.

Pauling, L. 1970. *General Chemistry*. Dover Publications, New York, 989 pp.

Pfannkuch, H. O. and T. C. Winter. 1984. Effect of anisotropy and groundwater system geometry on seepage through lakebeds, 1. Analog and dimension analysis. *J. Hydrol*. 75: 213–237.

Quinn, P., K. Beven, P. Chevallier, and O. Planchon. 1991. The prediction of hillslope flow paths for distributed hydrological modeling using digital terrain models. *Hydrol. Process.* 5: 59–79.

Richardson, J. L. 1997. Soil development and morphology in relation to shallow ground water—An interpretation tool. In K. W. Watson, and A. Zaporozec (Eds.) *Advances in Ground-Water Hydrology: A Decade of Progress*. American Institute of Hydrology, St. Paul, MN, pp. 229–233.

Richardson, J. L., J. L. Arndt, and J. Freeland. 1994. Wetland soils of the prairie potholes. *Adv. Agron*. 52: 121–171.

Richardson, J. L. and R. J. Bigler. 1984. Principle component analysis of prairie pothole soils in North Dakota. *Soil Sci. Soc. Am. J*. 48: 1350–1355.

Richardson, J. L. and F. D. Hole. 1978. Influence of vegetation on water repellency in selected western Wisconsin soils. *Soil Sci. Soc. Am. J*. 42: 465–467.

Richardson, J. L., L. P. Wilding, and R. B. Daniels. 1992. Recharge and discharge of ground water in aquic conditions illustrated with flownet analysis. *Geoderma* 53: 65–78.

Riveros-Iregui, D. A., R. E. Emanuel, D. J. Muth, B. L. McGlynn, H. E. Epstein, D. L. Welsch, V. J. Pacific, and J. M. Wraith. 2007. Hysteresis between soil temperature and soil CO_2 is controlled by soil water content. *Geophys. Res. Lett*. 34: L17404.

Rosenberry, D. O. and J. W. LaBaugh. 2008. *Field Techniques for Estimating Water Fluxes between Surface Water and Ground Water*. U.S. Geological Survey Techniques and Methods 4-D2.

Rosenberry, D. O. and T. C. Winter. 1997. Dynamics of water-table fluctuations in a upland between two prairie-pothole wetlands in North Dakota. *J. Hydrol*. 191: 266–289.

Schoeneberger, P. J., D. A. Wysocki, E. C. Benham, and W. D. Broderson. 1998. *Field Book for Describing and Sampling Soils*. National Soil Survey Center, Natural Resources Conservation Service, USDA, Lincoln, NE.

Schot, P. 1991. *Solute transport by groundwater flow to wetland ecosystems*. PhD thesis, University of Utrecht, Geografisch Instituut Rijksuniversiteirt, The Netherland 134 p.

Schwartzendruber, D., M. J. De Boodt, and D. Kirkham. 1954. Capillary intake rate of water and soil structure. *Soils Sci. Soc. Am. Proc.* 18: 1–7.

Simonson, R. W. 1959. Outline of a generalized theory of soil genesis. *Soil Sci. Soc. Am. Proc.* 23: 152–156.

Skaggs, R. L., D. Amatya, R. O. Evans, and J. E. Parsons. 1994. Characterization and evaluation of proposed hydrologic criteria for wetlands. *J. Soil Water Conserv.* 49(5): 501–510.

Skaggs, R. W., W. F. Hunt, G. M. Chescheir, and D. M. Amatya. 1995. Reference simulations for evaluating wetland hydrology. In K. L. Campbell (Ed.) *Versatility of Wetlands in the Agricultural Landscape*, Hyatt Regency, Tampa, Florida, September 17–20, 1995. Am. Soc. Agric. Engineers, St. Joseph, MI.

Skopp, J., M. D. Jawson, and J. W. Doran. 1990. Steady-state microbial activity as a function of soil water content. *Soil Sci. Soc. Am. J.* 54: 1619–1625.

Sloan, C. E. 1972. *Ground-Water Hydrology of Prairie Potholes in North Dakota*. U.S. Geol. Survey Professional Paper 585-C. U.S. Govt. Printing Office, Washington, DC.

Soil Survey Staff. 2014. *Keys to Soil Taxonomy*, 12th ed. USDA-Natural Resources Conservation Service, Washington, DC.

Sprecher, S. W. 2000. *Installing Monitoring Wells/Piezometers in Wetlands.* ERDC TN-WRAP-00-02. U.S. Army Engineer Research and Development Center, Vicksburg, MS.

Sprecher, S. W. and A. G. Warne. 2000. *Accesing and Using Meteorological Data to Evaluate Wetland Hydrology*. ERDC/EL TR-WRAP-00-1. U.S. Army Engineer Research and Development Center, Vicksburg, MS.

Steinwand, A. L. and J. L. Richardson. 1989. Gypsum occurrence in soils on the margin of semipermanent prairie pothole wetlands. *Soil Sci. Soc. Am. J.* 53: 836–842.

Stewart, R. E. and H. A. Kantrud. 1971. *Classification of Natural Ponds and Lakes in Glaciated Prairie Region. U.S. Fish Wildl. Serv., Res. Publ.* Vol. 92. U.S. Govt. Printing Office, Washington, DC.

Sumner, J. P, M. J. Vepraskas, and R. E. Kolka 2009. Methods to evaluate normal rainfall for short-term wetland hydrology assessment. *Wetlands* 29(3): 1049–1062.

Tischendorf, W. G. 1968. *Tracing stormflow to a varying source area in small forested watershed in the southeastern Piedmont*. PhD dissertation, Univ. of Georgia, Athens, GA.

Toth, J. 1963. A theoretical analysis of groundwater flow in small drainage basins. *J. Geophys. Res.* 68: 4197–4213.

U.S. Army COE. 1987. *Wetlands Delineation Manual*. Environmental Laboratory, U.S. Army Corps of Engineers Waterways Experiment Station, Vicksburg, MS.

U.S. Army COE. 2005. *Technical Standard for Water Table Monitoring of Potential Wetland Sites*. ERDC TN-WRAP-05-2. U.S. Army COE Research and Development Center, Visckburg, MS.

U.S. Army COE. 2010. Wetland hydrology indicators. In J. S. Wakley, R. W. Lichvar, and C. V. Noble (Eds.) *Regional Supplement to the Corps of Engineers Wetland Delineation Manual: Midwest Region*, Version 2.0. ERDC/EL TR-10-16. U.S. Army Engineer Research and Development Center, Vicksburg, MN.

U.S. Army COE St. Paul District. 1996. *U.S. Army COE St. Paul District, 17 April 1996, Guidelines for Submitting Wetland Delineations to the St. Paul District Corps of Engineers and Local Units of Government in the State of Minnesota. Public Notice 96-01078-SDE.* USA-COE, St. Paul, MN.

USDA NRCS. 1997. Hydrology tools for wetland determination. Chapter 19. In D. E. Woodward (Ed.) *Engineering Field Handbook*. USDA-NRCS, Fort Worth, TX.

USDA NRCS. 1994. National food security act manual, third edition (as amended). Washington, DC.

Vepraskas, M. J. and J. P. Williams. 1995. Hydraulic conductivity of saprolite as a function of sample dimensions and measurement technique. *Soil Sci. Soc. Am. J.* 59: 975–981.

Verones, F., D. Saner, S. Pfister, D. Baisero, C. Rondinini, and S. Hellweg. 2013. Effects of consumptive water use on biodiversity in wetlands of international importance. *Environ. Sci. Technol.* 47: 12248–12257.

Whittig, L. D. and P. Janitzky. 1963. Mechanisms of formation of sodium carbonate in soils I. Manifestations of biologic conversions. *J. Soil Sci.* 14: 322–333.

Winter, T. C. 1983. The interaction of lakes with variably saturated porous media. *Water Resour. Res.* 19: 1203–1218.

Winter, T. C. 1988. A conceptual framework for assessing cumulative impacts on hydrology of non-tidal wetlands. *Environ. Manag.* 12: 605–620.

Winter, T. C. 1989. Hydrologic studies of wetlands in the northern prairies. In A. Van der Valk (Ed.) *Northern Prairie Wetlands*. Iowa State University Press, Ames, IA, pp. 16–54.

Winter, T. C. 1992. A physiographic and climatic framework for hydrologic studies of wetlands. In Roberts, R. D. and M. L. Bothwell (Eds.) *Aquatic Ecosystems in Semi-arid Regions: Implications for Resource Management*. *N. H. R. I. Symposium Series*, Vol. 7. Environment Canada Saskatoon, Saskatchewan, Canada, pp. 127–148.

Winter, T. C., J. W. Harvey, O. L. Franke, and W. M. Alley. 1998. *Ground Water and Surface Water a Single Resource*. U.S. Geological Survey Circular 1139, Denver, CO.

Winter, T. C. and M.-K. Woo. 1990. Hydrology of lakes and wetlands. In M. G. Wolman, and H. C. Riggs (Eds.) *Surface Water Hydrology. The Geology of North America*, *v. 0-1*. Geological Society of America, Boulder, CO, pp. 159–187.

4

Redox Chemistry of Hydric Soils

Michael J. Vepraskas, Matthew Polizzotto, and Stephen P. Faulkner

CONTENTS

Introduction

Hydric soils are described in Chapter 2 as soils that form under anaerobic conditions that develop while the soils are inundated or saturated near their surface. These soils can form under a variety of hydrologic regimes that include nearly continuous saturation (swamps and marshes), short duration flooding (riparian systems), and periodic saturation by groundwater. The most significant effect of excess water is the isolation of the soil from the atmosphere and the slowing of O_2 from entering the soil. The blockage of atmospheric O_2 induces biological and chemical processes that change the soil from an aerobic and oxidized state to an anaerobic and reduced state. This shift in the aeration status of the soil allows chemical reactions to occur that develop the common characteristics of hydric soils such as the accumulation of organic carbon in A horizons, gray-colored subsoil horizons, and production of gases such as H_2S and CH_4. The creation of anaerobic conditions requires adaptations in plants if they are to survive in the anaerobic hydric soils. In addition, redox reactions in wetland soils help regulate environmental quality, impacting release and sequestration of greenhouse gases, nutrient pollution of surface water, and mobilization of potentially toxic trace elements to groundwater.

This chapter discusses the chemistry of hydric soils by focusing on the oxidation–reduction reactions that affect certain properties and functions of hydric soils and form the indicators by which hydric soils are identified (Chapter 7). Both the biological and chemical functions of wetlands are controlled to a large degree by oxidation–reduction chemical reactions (Mitsch and Gosselink 2007). The fundamentals behind these reactions will be reviewed in this chapter along with methods of monitoring these reactions in the field, and the effects of these reactions on environmental quality and major nutrient cycles in wetlands.

In our experience, soil chemistry is probably the subject least understood by students of hydric soils and wetlands in general. Therefore, the following treatment is intended to be simple, and to cover those topics that can be related to the field study of hydric soils. Students wishing more detailed treatments are encouraged to consult the work of Ponnamperuma (1972) in particular, as well as the discussion of redox reactions in McBride (1994) and Sparks (2003).

Oxidation and Reduction Basics

Oxidation–reduction (redox) reactions govern many of the chemical processes occurring in saturated soils and sediments (Baas-Becking et al. 1960). Redox reactions transfer electrons among atoms. As a result of the electron transfer, electron donor atoms increase in valence to become more positively charged, and the electron acceptor atoms decrease in valence and become more negatively charged. Such changes in valence usually alter the phase in which the atom occurs in the soils such as causing solid minerals to dissolve or dissolved ions to turn to gases. The loss of one or more electrons from an atom is known as *oxidation* because in the early days of chemistry the known oxidation reactions such as rust formation, always involved oxygen. The gain of one or more electrons by an atom is called *reduction* because the addition of negatively charged electrons reduces the overall valence of the atom. Each complete redox reaction contains an oxidation and a reduction component that are called *half-reactions*. Redox reactions are more easily understood and evaluated when the oxidation and reduction half-reactions are considered separately. This is appropriate because oxidation and reduction processes each produce different effects on the soil.

For example, in aerobic soils organic compounds such as the carbohydrate glucose can be oxidized to CO_2 as shown in the following reaction:

$$C_6H_{12}O_6 + 6O_2 \rightarrow 6CO_2 + 6H_2O \tag{4.1}$$

This reaction can be broken down into an oxidation half-reaction and a reduction half-reaction.

$$C_6H_{12}O_6 + 6H_2O \rightarrow 6CO_2 + 24e^- + 24H^+ \text{ (Oxidation)} \tag{4.2}$$

$$6O_2 + 24e^- + 24H^+ \rightarrow 12H_2O \text{ (Reduction)} \tag{4.3}$$

The basic oxidation half-reactions in soils are catalyzed by microorganisms during their respiration process (Chapter 5). The respiration is responsible for releasing one or more electrons, as well as hydrogen ions. Oxidation occurs whenever heterotrophic microorganisms are using organic tissues as their carbon source for respiration, as when organic tissues are being decomposed in soils. For this discussion, bacteria will be considered the major group of organisms initiating the oxidation processes in soil. Organic tissues are the major source of electrons, and when the tissues are oxidized the electrons released are used for reducing reactions. The most important point to remember is that when organic tissues are not present, or when bacteria are not respiring, redox reactions of the type discussed in this chapter will not occur in the soil.

Alternate Electron Acceptors

Electron acceptors are the substances reduced in the redox reactions. Oxygen is the major electron acceptor used in redox reactions in aerobic soils. However, in anaerobic soils, where O_2 is not present, other electron acceptors have to be used by bacteria if they are to continue their respiration by oxidizing organic compounds. The major electron acceptors that are available in anaerobic soils are contained in the following compounds: NO_3^-, MnO_2, $Fe(OH)_3$, SO_4^{2-}, and CO_2 (Ponnamperuma 1972, Turner and Patrick 1968).

Theoretically, the electron acceptors are reduced in anaerobic soils in the order shown above due to the thermodynamics of the different reactions (Sparks 2003). In an idealized case, when organic compounds are being oxidized, O_2 will be the only electron acceptor used while it is available. When the soil becomes anaerobic upon the complete reduction of most available O_2, then NO_3^- will be the acceptor reduced while it is available. This same sequence is followed by the other compounds shown. Thus, if O_2 is never depleted, the reduction of the other compounds will never occur. While not all bacteria use the same electron acceptors, we will assume that most soils contain all microbial species necessary to reduce each of the electron acceptors noted earlier.

The order of reduction discussed above is idealized and probably does not occur in soil horizons exactly as predicted from theoretical grounds. It has been observed that the reduction of Fe^{3+} and Mn^{4+} can occur in a soil, even though, some O_2 is still present (McBride 1994). The theoretical order of reduction requires that the soil's *Eh* value—its redox potential—be an equilibrium value such that all redox half-reactions have adjusted to it. For this to happen, the soil's *Eh* must remain stable over a certain time period, be the same across the horizon, and all-electron acceptors have to be able to react at a similar rate. A soil's *Eh* is never stable for long if the soil is affected by a fluctuating water table. Furthermore, *Eh* values will vary across a soil horizon at some periods because organic tissues are not uniformly distributed: roots can be found at cracks or in large channels, but not in some parts of the soil matrix. This means that reducing reactions that are occurring around a dead root will not be the same as those occurring in an air bubble a few centimeters away. In addition, electron acceptors also do not become reduced at similar rates.

A discussion of reaction kinetics is beyond the scope of this chapter, but the topic has been reviewed by McBride (1994), who provides a thorough discussion of the order of reduction of the electron acceptors. Despite these inherent problems, the general order of reduction presented above is useful for understanding the general reduction sequence that occurs in hydric soils.

Principal Reducing Reactions in Hydric Soils

Reducing reactions, especially those that use compounds other than O_2, are the ones most responsible for the distinctive chemical processes that occur in hydric soils such as denitrification, production of mottled soil colors, and production of hydrogen sulfide and methane gases. Common reducing reactions found in hydric soils are listed in Table 4.1.

Because the electron acceptors most commonly used are compounds that contain oxygen, the basic reducing reactions produce water as a by-product, as shown in Equation 4.3 and Table 4.1. This process removes H^+ ions from solution and causes the pH of acid soils to rise during the reduction process.

Oxygen reduction occurs when organic tissues (organic matter) are being oxidized in a soil horizon that lies above the water table and in a soil that is not covered by water. Oxygen reduction can also occur in saturated soils where O_2 is dissolved in the soil solution. This frequently occurs when water (rainfall) has recently infiltrated a soil. When oxygen reduction has removed virtually all dissolved O_2, organic tissues decompose more slowly. If anaerobic conditions and slow decomposition are maintained for a long period, then organic C accumulates and organic soils may form (Chapter 10).

Denitrification is the reduction of nitrate to dinitrogen gas by the following reaction:

$$2NO_3^- + 10e^- + 12H^+ \rightarrow N_2 + 6H_2O \tag{4.4}$$

Other gaseous by-products containing N are also possible (Firestone and Davidson, 1989). The reaction is similar to oxygen reduction in that water is produced. This reaction improves water quality by removing NO_3^-, but it has no direct impact on soil properties such as color or organic C content, which can be used to identify hydric soils in the field.

TABLE 4.1

Half-Cell Reducing Reactions and the Equations Used to Calculate the Phase Change Lines Shown in Figure 4.1

Half-Cell Reaction	**Redox Potential (*Eh*, mV)**
$1/4O_2 + H^+ + e^- = 1/2H_2O$	$1229 + 59\log(P_{O_2})^{1/4} - 59\,pH$
$1/5NO_3^- + 6/5H^+ + e^- = 1/10N_2 + 3/5H_2$	$1245 - 59[\log(P_{N_2})^{1/10} - \log(NO_3)^{1/5}] - 71\,pH$
$1/2MnO_2 + 2H^+ + e^- = 1/2Mn^{2+} + H_2O$	$1224 - 59\log(Mn^{2+}) - 118\,pH$
$Fe(OH)_3 + 3H^+ + e^- = Fe^{2+} + 3H_2O$	$1057 - 59\log(Fe^{2+}) - 177\,pH$
$FeOOH + 3H^+ + e^- = Fe^{2+} + 2H_2O$	$724 - 59\log(Fe^{2+}) - 177\,pH$
$1/2Fe_2O_3 + 3H^+ + e^- = Fe^{2+} + 3/2H_2O$	$707 - 59\log(Fe^{2+}) - 177\,pH$
$1/8SO_4^{2-} + 5/4H^+ + e^- = 1/8H_2S + 1/2H_2O$	$303 - 59[\log(P_{H_2S})^{1/8} - \log(SO_4{}^{2-})^{1/8}] - 74\,pH$
$1/8CO_2 + H^+ + e^- = 1/8CH_4 + 1/4H_2O$	$169 - 59[\log(P_{CH_4}) - \log(P_{CO_2})^{1/2}] - 59\,pH$
$H^+ + e^- = 1/2H_2$	$0.00 - 59[\log(P_{H_2})^{1/2} - 59\,pH$

Manganese reduction occurs after most of the nitrate has been reduced, converting manganese from the 4+ into the 2+ valence states.

$$MnO_2 + 2e^- + 4H^+ \rightarrow Mn^{2+} + 2H_2O \tag{4.5}$$

The MnO_2 is a mineral with a black color. When reduced, the oxide dissolves and Mn^{2+} stays in solution and can move with the soil water.

Iron reduction is the reducing reaction occurring in hydric soils that most greatly affects soil color. Iron behaves much like Mn and has two oxidation states—2+ and 3+. When oxidized, the ferric form of Fe (Fe^{3+}) occurs as an oxide or hydroxide mineral. All of these oxidized forms of Fe impart brown, red, or yellow colors to the soil. The reduced ferrous Fe (Fe^{2+}) is colorless, soluble, and can move through the soil. The reducing reaction that ferric Fe undergoes varies with the type of ferric-Fe mineral present, as shown in Table 4.1. For amorphous Fe minerals, the reducing reaction is

$$Fe(OH)_3 + e^- + 3H^+ \rightarrow Fe^{2+} + 3H_2O \tag{4.6}$$

Sulfate reduction is performed by obligate anaerobic bacteria (Germida 1998). The basic reaction is similar to that for nitrate reduction, and it too produces a gaseous product.

$$SO_4^{2-} + 8e^- + 10H^+ \rightarrow H_2S + 4H_2O \tag{4.7}$$

The H_2S gas has a smell like that of rotten eggs. It can be easily detected in the field, but occurs most often near coasts where seawater supplies SO_4^{2-} for reduction.

Carbon dioxide reduction produces methane, the major component in the natural gas used in homes. This reaction is also similar to the others that produce a gaseous by-product.

$$CO_2 + 8e^- + 8H^+ \rightarrow CH_4 + 2H_2O \tag{4.8}$$

Methane is an inflammable gas. It can be identified in the field when it is collected in water-filled plastic bags that are inverted and placed on the surface of a submerged soil for 24 h. If the bubble of gas trapped in the bag is allowed to escape through a pinhole placed in the bag and if it ignites in the presence of a flame, it is assumed to be methane (J. M. Kimble, USDA, personal communication).

Reduction of minor and trace chemical species in hydric soils may also occur, in addition to the primary reducing reactions described above. For instance, many potentially toxic trace elements—including arsenic, chromium, and uranium—may exist in a variety of valence states (Sparks 2003), and their occurrence may be naturally or anthropogenically derived. The timing and location of reduction reactions involving these minor and trace species are based on the soil *Eh* and energetics of the particular reactions. Redox transformations can alter the mobilities and toxicities of these species, potentially impacting water quality and plant health, but, in general, the specific species transformations must be determined analytically rather than through visual observation.

Factors Leading to Reduction in Soils

Four conditions are needed for a soil to become anaerobic and to support the reducing reactions discussed above (Meek et al. 1968, Bouma 1983): (1) the soil must be saturated or inundated to exclude atmospheric O_2; (2) the soil must contain accessible organic tissues (organic matter) that can be oxidized or decomposed; (3) a microbial population must be respiring and oxidizing the organic tissues; and (4) the water should be stagnant or moving very slowly. Saturation or inundation is needed to keep the atmospheric O_2 out of the soil. Exclusion of atmospheric O_2 is probably the major factor that determines when reduction can occur in the soil. The presence of oxidizable organic tissues is probably the most important factor determining whether or not reduction occurs in a saturated soil (Beauchamp et al. 1989). Some soils are known to be saturated yet do not display any signs that reducing reactions such as Fe^{3+} reduction have occurred. In most instances, such soils simply lack the oxidizable organic tissues needed to supply the electrons used in reducing reactions (Couto et al. 1985).

A respiring microbial population is essential to the formation of reduced soils. Bacteria are widespread, abundant, varied and adapted to function in the climates in which they occur. As reducing chemical reactions are studied more extensively in the field, it is becoming clear that they occur more frequently than originally thought (Megonigal et al. 1996, Clark and Ping 1997). Lastly, stagnant or near-stagnant water is needed for reducing reactions to occur (Gilman 1994). Fast-moving water, particularly at the surface, retards the onset of reduction by supplying oxygen to the soil. While the water is in motion, its O_2 is difficult to deplete.

Quantifying Redox Reactions in Soils

Thermodynamic Principles

Oxidation–reduction reactions can be expressed thermodynamically using the concept of redox potential (*Eh*). This discussion begins with a review of thermodynamic principles that can be applied directly in the field to evaluate which primary redox reactions are occurring in a soil. The theory behind redox potential can be derived by considering the general reducing equation.

$$\text{Oxidized molecule} + m\text{H}^+ + n\text{ electrons} = \text{Reduced molecule} \tag{4.9}$$

where m is the number of moles of protons, and n is the number of moles of electrons used in the reaction. This reaction can be expressed quantitatively by calculating the Gibbs free energy (ΔG) for the reaction.

$$\Delta G = \Delta G^\circ + RT\ln\frac{(\text{Red})}{(\text{Ox})(\text{H}^+)^m} \tag{4.10}$$

where ΔG° is the standard free energy change, R is the gas constant, T is absolute temperature, and (Red) and (Ox) represent the activities of reduced and oxidized species. This equation can be transformed into one more applicable to us by converting the Gibbs free energy into a unit of voltage using the relationship $\Delta G = -nEF$.

$$Eh = E^\circ - \frac{RT}{nF}\ln\frac{(\text{Red})}{(\text{Ox})} - \frac{mRT}{nF}\ln(H^+) \tag{4.11}$$

where Eh is the electrode potential (redox potential) for the reaction, E° is the potential of the half-reaction under standard conditions (unit activities of reactants under 1 atmosphere of pressure and a temperature of 298°K), and *F* is the Faraday constant. Equation 4.11 is called the Nernst equation. Substituting values for *R*, *F*, and *T* of 8.3 J/K mol, 9.65×10^4 C mol^{-1}, and 298°K, respectively, converting the logarithm, and substituting pH for $-\log(H^+)$ the Nernst equation can be simplified to

$$Eh(\text{mV}) = E^\circ - \frac{59}{n}\log\frac{(\text{Red})}{(\text{Ox})} + \frac{59\,m}{n}\text{pH} \tag{4.12}$$

The Nernst equation shows that the reduction of an element will create a specific *Eh* value at equilibrium; however, the exact *Eh* value will vary with soil pH and the concentration (activity) of oxidized and reduced species in the soil. This equation has practical value for monitoring the development of reducing conditions in hydric soils in the field.

Eh/pH Phase Diagrams

Equation 4.12 is used in Figure 4.1 to portray graphically the major reducing reactions occurring in hydric soils. The figure was prepared using the equations shown in Table 4.1, which were modified from the half-reactions described earlier. The equations represent the following conditions: dissolved species were assumed to have activities of 10^{-5} M, partial pressures for O_2 and CO_2 were 0.2 and 0.8 atm, respectively, and partial pressures of the remaining gases were assumed to be 0.001 atm, which approximate what might be found in nature (McBride 1994).

The upper and lower lines in Figure 4.1 are the theoretical limits expected for redox potentials in soils because of the buffering effect of water on redox reactions. *Eh* values, above the upper line shown in Figure 4.1, are prevented at equilibrium because water in the soil would oxidize to O_2 and supply electrons that would lower the *Eh*. *Eh* values

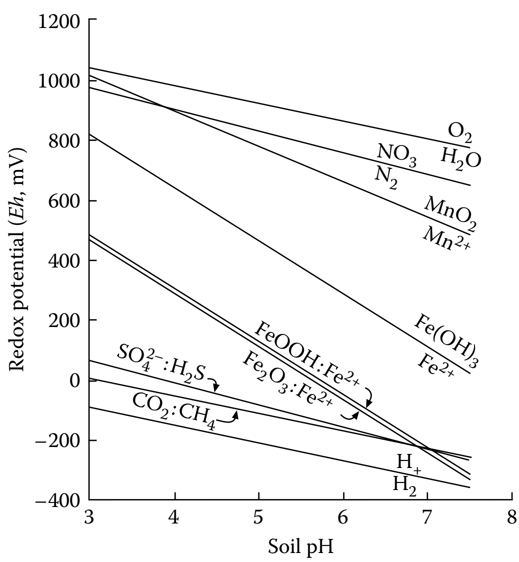


FIGURE 4.1
An *Eh*–pH phase diagram for the reducing reactions shown in Table 4.1. The lines were computed for the following conditions: dissolved species were assumed to have an activity of 10^{-5} M, partial pressures for O_2 and CO_2 were 0.2 and 0.8 atm, respectively, and partial pressures of the remaining gases were assumed to be 0.001 atm.

below the lower line are prevented because water (which supplies H^+) would be reduced to H_2, consuming electrons and raising the *Eh*. The *Eh* values at which the other reducing reactions occur vary with pH, and also vary with the assumptions regarding the concentrations noted earlier. These theoretical limits vary with pH as described by the Nernst equation.

The order or sequence for which the electron acceptors are reduced is clearly shown in Figure 4.1. The sequence changes somewhat for different pHs. The Fe oxides shown in Table 4.1, each has separate phase lines. The nearly amorphous $Fe(OH)_3$ minerals (ferrihydrite) reduce at a higher *Eh* value for a given pH than do the crystalline minerals of FeOOH (goethite) or Fe_2O_3 (hematite). Field studies have shown that the $Fe(OH)_3$ minerals occupy 30%–60% of these Fe minerals in hydric soils (Richardson and Hole 1979).

Reliability of Phase Diagrams for Field Use

Eh/pH phase diagrams are useful for showing how reduction and oxidation of a given species vary with the pH of the solution, and they also show the relationship among the different elements that undergo redox reactions. Once a redox phase diagram is in hand, the next logical step is to measure *Eh* and pH in the field and use these data to predict the phase a given element is in. It is possible to do this for some redox reactants, but phase diagrams have two potential problems that directly limit field applications. The first deals with mixed redox couples, and the second with the kinetics of redox reactions.

Mixed Redox Couples

The lines on an *Eh*/pH diagram show the *Eh* and pH values where a specific redox couple (half-reaction) is expected to undergo a phase change and attain the concentration that was used to develop the diagram. Each line on the phase diagram was computed by assuming that both the *Eh* and pH values measured in the soil solutions were influenced only by a single redox half-reaction and that equilibrium had been achieved. This will generally not be the case if other substances, which are also undergoing redox reactions, are present in the soil solution, and if the soil's *Eh* value is changing over time. In such cases, the soil's *Eh* value would be a *mixed potential*, or an average potential determined by a number of the half-reactions shown in Table 4.1, and not simply the result of a single redox half-reaction. These "average *Eh* values" complicate the use of phase diagrams for interpretations of redox data because they are not in equilibrium with each other, and, therefore, the actual *Eh* at which a phase change will occur cannot be predicted precisely using the equations of Table 4.1.

The presence of mixed redox potentials also creates problems when attempting to adjust *Eh* values for different pHs. For example, where the ratio of protons to electrons (m/n in Equation 4.12) is unity in the half-cell reaction, the Nernst equation predicts a 59 mV change in *Eh* per pH unit. This value is sometimes used to adjust measured redox potentials for comparison at a given pH, but as shown in Table 4.1, the ratio of m/n varies for different redox couples and ranges from –59 to –177 mV/pH unit. The *Eh*/pH slope predicted from the Nernst equation assumes that a specific redox couple controls the pH of the system. While this may be true for controlled laboratory solutions, the pH of natural soils and sediments is buffered by silicates, carbonates, and insoluble oxide and hydroxide minerals which are not always involved in redox reactions (Bohn et al. 1985, Lindsay 1979). Therefore, it is not surprising that measured slopes in natural soils deviate from the predicted values. Applying a theoretical correction factor to adjust *Eh* values for pH differences among soils may be inappropriate for natural conditions (Bohn 1985, Ponnamperuma 1972). We recommend that *Eh* values measured in soils not be adjusted to a common pH, but rather that the pH of the soil be measured and reported whenever *Eh* values are reported.

Mixed redox couples can also alter the apparent slopes of the phase lines shown in Figure 4.1. For instance, a change of +177 mV per pH unit is the predicted slope for the reduction of $Fe(OH)_3$ to Fe^{2+} (Table 4.1) based on the m/n value of 3 (i.e., 24/8). In a series of experiments where he added different kinds of plant organic matter to several different kinds of soils, Zhi-guang (1985) found that this slope varied as a function of the ratio of ferrous iron to organic matter. In sandy soils with almost no Fe^{2+}, the slope matched the theoretical value of 59 mV/pH. As the Fe^{2+} concentration increased, the slope also increased but did not reach the theoretical value of –177 mV/pH. On the other hand, Collins and Buol (1970) found good agreement between the measured and theoretical *Eh*/pH relationship for soils containing more Fe minerals. In summary, we feel phase diagrams such as those shown in Figure 4.1 will be most useful for interpreting redox data for elements that are abundant (e.g., Fe) in a soil, and where soil pHs are influenced by the redox reactions and are not buffered by carbonates as would be expected at soil pHs >7.

Reaction Kinetics

Another problem that complicates the use of phase diagrams with natural *Eh* data is that some redox reactions occur much more slowly than others. This is particularly true for the reduction of O_2, and MnO_2 (McBride 1994). The effect of this is that the actual *Eh* at which detectable amounts of reduced species of these compounds occurs tends to be 200–300 mV lower than what would be predicted in Figure 4.1. This means that redox potential measurements in soils may not relate well to the chemical composition of soil solutions that are predicted by Figure 4.1. On the other hand, redox reactions related to Fe have been found to begin in soils (using Pt electrodes) near the *Eh* values specified in Figure 4.1. Reduction of SO_4 and CO_2 also begin at *Eh* values similar to those predicted in Figure 4.1. In summary, phase diagrams can be useful to interpret data for transformation of specific Fe minerals in soils, but caution is needed for predicting when reduction occurs for O_2, NO_3^-, and MnO_2.

The Concept of *pe*

Redox reactions written as half-reactions treat electrons (e^-) as a reactive species very similar to H^+. While free electrons do not occur in solution in any appreciable amount, the electrons can be considered as having a specific activity. Analogous to the concept of pH, electron activity is expressed as *pe*, which has been defined as (Ponnamperuma 1972)

$$pe = -\log(e^-) = \frac{Eh(\text{mV})}{59} \tag{4.13}$$

Solutions with a high electron activity (low *pe*) and low *Eh* value conceptually have an abundance of "free electrons." These solutions are expected to reduce O_2, NO_3, MnO_2, etc. Solutions that have a low electron activity (high *pe*) and high *Eh* value can be thought of as having virtually no "free electrons," and will maintain the elements of O, N, Mn, Fe, etc., in their oxidized forms. The *pe* can also be used as a substitute for *Eh* in Equation 4.12.

$$pe = \frac{E^\circ}{59} - \frac{1}{n}\log\frac{(\text{Red})}{(\text{Ox})} - \frac{m\,\text{pH}}{n} \tag{4.14}$$

This equation can be used to develop phase diagrams like that shown in Figure 4.1. Although the *pe* concept is useful for chemical equilibria studies, it is a theoretical concept that cannot be measured directly in nature. We will continue to use redox potential (*Eh*) as our measure of reducing intensity because this voltage can be measured in the field.

Measuring Reduction in Soils

Chemical Analyses

The chemistry of hydric soils can be evaluated in a general sense by measuring the concentrations of reduced species in solution. If, for example, there is no measurable O_2 in solution, the soil is known to be anaerobic. If Fe^{2+} is detected in solution, we can predict from theoretical grounds that the soil is probably anaerobic, that denitrification has occurred (if NO_3^- was present initially), that manganese reduction has taken place, but that the reduction of SO_4^{2-} and CO_2 may or may not have occurred. Reaction kinetics and microsite reduction can create exceptions to these interpretations. An additional complicating factor is that expected reduction reaction products may be scavenged from the soil solution, for instance through secondary mineral precipitation, limiting their direct measurement and interpretation of ongoing redox processes. Chemical evaluations of all reduced species in solution are expensive and usually used only for research purposes as described in the "Nutrient Pools, Transformations, and Cycles" section of this chapter.

Dyes

A less expensive alternative to measuring soil solution chemistry is to use a dye that reacts with reduced forms of key elements. The most widely used dyes for field evaluations of reduction react with Fe^{2+}. Childs (1981) discussed the use of α,α′-dipyridyl in the field. Heaney and Davison (1977) showed that the α,α′-dipyridyl reagent reliably distinguished Fe^{2+} from Fe^{3+}, and that dye results corresponded well with measurements of the concentration of these species. Other dyes, such as 1,10-phenanthroline, are available to detect Fe^{2+} in reduced soils, and all can be used in similar ways (Richardson and Hole 1979). Dyes work quickly in the field and are easy to use. To test for Fe^{2+} in the field, a sample of *saturated* soil is extracted, or a soil pit is excavated and the dye solution immediately sprayed onto an exposed soil horizon. If Fe^{2+} is present, it will react with the dye within one minute and change color. Both 1,10-phenanthroline and α,α′-dipyridyl turn red when they react with Fe^{2+}. It must be remembered that these dyes detect only Fe^{2+}. If a positive reaction occurs after the dye is applied to a soil sample, it can be assumed that the soil is reduced in terms of Fe and that the soil must also be anaerobic. If no reaction to the dye is found, then all we know is that Fe^{2+} is not present. The soil, in this case, may be anaerobic, but not Fe-reduced, or it may be aerobic. Either of these two cases will produce a negative reaction to the dye solution.

A 0.2% solution of α,α′-dipyridyl dye is used in the field by soil classifiers of the USDA Natural Resources Conservation Service (Soil Survey Staff 1999). It is prepared by first dissolving 77 g of ammonium acetate in 1 L of distilled water. Then 2 g of α,α′-dipyridyl dye powder is added, and the mixture stirred until the dye dissolves. The dye powder and solution are both sensitive to light and should be kept in brown bottles or in the dark. This solution can be applied with a dropper to freshly broken surfaces of saturated soils. If a pink (low ferrous iron) or red (high ferrous iron) color develops within a minute, ferrous iron is present. This procedure uses a neutral (pH ~ 7.0) solution, which avoids potential errors associated with photochemical reduction of ferric–organic complexes. Avoid spraying onto soils contacted by steel augers or shovels, because these may give false-positive tests. For darkcolored soils (e.g., Mollisols and Histosols), the use of white filter paper improves the ability to observe color development.

False-positive errors from photochemical reduction of ferric–organic compounds can occur when samples to which the dye has been applied are exposed to bright sunlight. In addition, exposure to air can rapidly oxidize Fe^{2+}–Fe^{3+} when pH >6 (Theis and Singer 1973) and produce a false-negative result. Childs (1981) describes the development of the test and the errors associated with the photochemical reduction of ferric–organic complexes.

IRIS Tubes

Another technique that can be used to assess reduction in soils uses IRIS tubes. IRIS (an indicator of reduction in soils) are polyvinyl chloride (PVC) tubes that have been coated with a Fe oxyhydroxide paint (Castenson and Rabenhorst 2006, Jenkinson and Franzmeier 2006). When placed in a reduced soil, the Fe paint dissolves exposing the underlying white color of the PVC tube. A recommended procedure for using IRIS tubes to identify hydric soils was presented by Rabenhorst (2008, 2012). IRIS tubes are inserted into the soil so that the coated portion of the tube extends from the surface to a depth of 30 cm. Replicated tubes at one location are recommended. The tubes can be checked after one month or so to determine if Fe reduction has occurred in the soil. The proportion of a horizon that has been reduced can also be estimated from the amount of Fe paint that has been removed. IRIS tubes have proven to be a simple and reliable method for assessing reduction in soils. They collect data passively while in the soil in that the scientist collects the data when the tubes are removed from the soil. IRIS tubes have the added advantage in that they can be placed in remote locations, where regular weekly site visits may be impractical, and used to collect data over long (i.e., 6 month) time periods.

Redox Potential Measurements

Redox potential (*Eh* in Equation 4.12) is a voltage that can be measured in the soil and used to predict the types of reduced species that would be expected in the soil solution. The *Eh* measurements are evaluated along with soil pH data and an *Eh*/pH phase diagram such as that shown in Figure 4.1. The redox potential voltage must be measured between a Pt-tipped electrode and a reference electrode that creates a standard set of conditions. Platinum electrodes are sometimes called microelectrodes because they consist of a small piece of Pt wire that is placed in the soil. The Pt wire is assumed to be chemically inert and only conducts electrons. It generally does not react itself with other soil constituents and does not oxidize readily as do Fe, Cu, and Al metals. Reduced soils transfer electrons to the Pt electrode while oxidized soils tend to take electrons from the electrode. For actual redox potential measurements, the electron flow is prevented. The potential or voltage developed between the soil solution and a reference electrode is measured with a meter that has been designed to detect small voltages. The voltages developed in soil range from approximately +1 to –1 V, and are usually expressed in millivolts (mV).

There are several methods of Pt electrode construction, but they all follow the same basic design (Faulkner et al. 1989, Patrick et al. 1996). For soil systems, 18-gauge platinum wire (approximately 1 mm in diameter) is preferred because it is more resistant to bending when inserted in the soil. The Pt wire is cut into 1.3-cm segments, with wire-cutting pliers that are used only for cutting platinum and cleansed in a 1:1 mixture of concentrated nitric and hydrochloric acids for at least 4 h. This removes any surface contamination that could occur during cutting or handling. The cut wire segments are then soaked overnight in distilled, deionized water.

For field studies of less than 3 years duration, welding or fusing the platinum directly to a 12- or 14-gauge copper wire or brass rod is the least complicated method to use. All exposed metal except Pt must be insulated with a nonconducting material (e.g., heat-shrink tubing) and a waterproof epoxy. This welded/fused design is most appropriate for studies of less than 3 years because many epoxy cements are not stable for extended periods under continuous exposure to water. Longer measurement periods are better served by a glass body electrode (see Patrick et al. 1996 for a complete description).

Platinum electrodes can be "permanently" installed in the soil and left in place for up to a year to monitor a complete wetting and drying cycle. After a year, some electrodes should be removed and retested in the laboratory to ensure that problems related to component breakdown are not occurring. The installation process must seal the electrodes from the movement of air or water from the surface to the tip. This can be done by augering a hole, filling it with a slurry made from the extracted soil, and inserting the cleaned Pt electrode to the appropriate depth. The slurry must have the same chemical properties as the soil the Pt tip is placed in.

Redox potential measurements are made in the field using a portable pH/mV meter and a saturated calomel or silver/silver chloride reference electrode. Commercial voltmeters can be used, but not all of them register millivolts. The reference electrode normally is not permanently installed at the site. To begin readings, the reference electrode is pushed a short distance into wet or moist soil at the surface to ensure a good electrical contact. If the soil is relatively dry, a knife or soil probe is used to excavate a shallow hole to hold the electrode upright. Water should be poured into the hole to provide good electrical contact between the reference electrode and soil solution. If the soil is dry, a dilute salt solution (i.e., 5 g KCl in 100 ml H_2O) can be used to moisten the reference electrode hole and prevent a junction potential from being established between the reference electrode and the soil. The reference electrode is connected to the "common" terminal on the commercial meters. The other terminal (for voltage) is connected to a single Pt electrode that is buried in the soil. To take a measurement after the electrodes are connected to the meter, the meter is turned on, and the voltage recorded. It is best to not wait more than a few seconds to record the voltage. Voltages that are found to drift or not stabilize can be caused by meters with a low internal resistance that allows to current to flow through the meter (Rabenhorst et al. 2007).

Correcting Field Voltages to the Standard Hydrogen Electrode

The voltage measured in the field between the buried Pt wire and a reference electrode is not the redox potential or *Eh*. True redox potentials are measured against a standard hydrogen electrode that consists of a Pt plate with H_2 gas moving across its surface. Such an electrode is impractical for field use. Correction factors are used to adjust the field voltage measured with one type of reference electrode to the voltage that would have been measured had a standard hydrogen electrode been used. The correction

TABLE 4.2

Correction Factors Needed to Adjust Voltages Measured in the Field to Redox Potentials (*Eh*'s) for Two Commonly Used Reference Electrodes

	mV	
Temperature (°C)	**Calomel (Hg-Containing)**	**Ag/AgCl**
25	244	197
20	248	200
15	251	204
10	254	207
5	257	210
0	260	214

Note: The factors are added to field-measured voltages to correct the values to voltages measured with standard hydrogen electrodes. Correction factors for the Ag/AgCl electrode assume the electrode is filled with a saturated KCl solution.

factors for two common reference electrodes are listed in Table 4.2. The correction is simply

$$\text{Field Voltage} + \text{Correction Factor} = \text{Redox Potential}\,(Eh) \tag{4.15}$$

Temporal Variability in Redox Potential

Redox potential measurements made at a single point in the soil may change over the course of a year by 1000 mV or more if the soil is periodically saturated or flooded and reducing reactions occur. Less variation is expected in soils that never saturate as well as ones that are permanently inundated. An example of the variation in redox potential for one hydric soil is shown in Figure 4.2, where data for the mean of five redox potential measurements are plotted, along with the minimum and maximum values found for the same depth. Before the soil became saturated in 1998, the redox potential was above 600 mV, and the range in values among the five electrodes was about 100 mV, which is relatively small. Within a few days of the soil saturating due to a rising water table, the redox potential fell, but the rate of fall was not the same among all five electrodes. During the period of decrease in redox potential across the horizon, the range in values was over 600 mV. By day 60 (in 1998) the range in redox potentials again was approximately 100 mV, even though the mean potential was near 0 mV. Later periods of greater redox potential variability were associated with periodic draining and resaturation.

The type of variability illustrated in Figure 4.2 is real and must be expected when making redox potential measurements in hydric soils that undergo periodic saturation and drainage. The variability seems to be caused by the oxidation of organic tissues and the corresponding reducing reactions occurring in microsites (Crozier et al. 1995, Parkin 1987). Microsites are simply small volumes of soil on the order of 1–5 cm^3 that surround decomposing tissues such as a dead root or leaf. Examples of microsites where reduction occurred are described and illustrated in Chapter 7 (e.g., Figure 7.8). When the redox potentials shown in Figure 4.2 were >600 mV, the soil was unsaturated, and O_2 was controlling or poisoning the system. After saturation had occurred, the oxidation of organic tissues by bacteria continued. After dissolved O_2 had been depleted, alternate electron

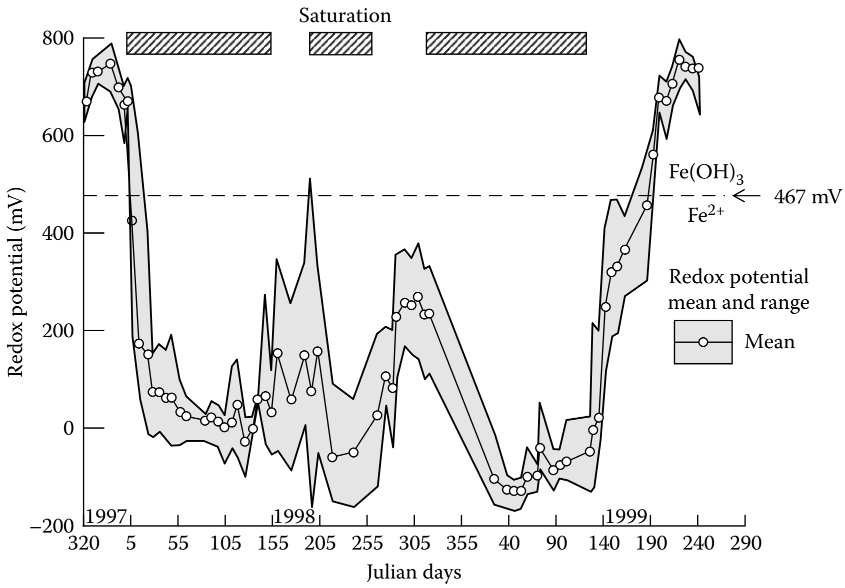


FIGURE 4.2
Variation in redox potential for a hydric soil at a depth of 30 cm. Data are the mean and range of five Pt electrodes. Variation among electrodes is greatest during periods when the soil is either saturating or draining, and less variation occurs when the soil is either saturated or drained for several weeks. Reduction of $Fe(OH)_3$ occurs within weeks of the soil saturating, and reduced Fe can be maintained even during intermittent periods when the soil is unsaturated.

acceptors were used in the reducing reactions. The Pt electrode that recorded the fastest drop in redox potential following saturation may have been adjacent to the decomposing tissue (near the microsite of reduction), while the electrode that responded most slowly may have been farther away. Although there are broad ranges in *Eh* following saturation due to the reduction occurring in microsites, over time the range in *Eh* narrows as the dissolved O_2 in the soil solution is depleted and a greater volume of soil becomes reduced.

To characterize the redox potentials in hydric soils, an adequate number of measurements must be made across the horizon to account for the variability expected in the redox potentials. Statistical analyses applied to redox data have usually indicated that 10 or more electrodes per depth are needed for an acceptable level of precision over a complete wetting/drying cycle. This is generally too expensive for routine use. We recommend, however, that at least five Pt electrodes be installed at each depth for which redox potential measurements are desired. Under no circumstances that we can imagine, should a single redox potential measurement be used to assess reducing conditions in the field.

In summary, soil redox potential measurements remain the most versatile tool we currently have for assessing reducing reactions economically for virtually any soil. The method, when properly applied, provides useful data on reducing reactions. The spatial and temporal variability in *Eh* is magnified during the initial periods of flooding/saturation and draining as the system changes from aerobic to anaerobic and back again. Because of these conditions, it is important to collect data over a period that includes a saturating and draining cycle. The most effective way to partially overcome the problem of spatial heterogeneity of a given soil is through replication of the measurement equipment.

Spatial Variability in Redox Processes

In addition to exhibiting temporal variability, as described above, the dominant redox processes within hydric soils may also vary substantially over space if the degree of soil saturation is also variable. For instance, redox potential changes that occurred over time in a landscape consisting of a hydric soil, transition zone, and upland area are shown in Figure 4.3. These redox potential data are the mean of five electrodes at a depth of 30 cm. The soils all had a pH of 5.0, and the $Fe(OH)_3$ phase line has been added to the figure. The occurrence of saturation clearly controls the fluctuation in redox potential among the three landscape positions. The upland soil never became saturated during the study period, and it can be seen that its redox potential remained high and fairly constant. The transitional soil was saturated for short periods (data not shown), but the redox potential never fell to a point where Fe reduction would have been expected. On the other hand, the hydric soil was saturated for an extended period, and Fe reducing conditions occurred for approximately 150 days.

Within hydric soils, it is common to observe a transitional sequence of reduction reactions through a vertical profile, particularly where the soil is not permanently submerged. Reduction reactions with depth in soils proceed according to the thermodynamics of their redox potentials (Table 4.1), with oxygen reduction occurring at the surface if aerobic conditions are present. As oxygen is depleted, denitrification, manganese reduction, iron reduction, sulfate reduction, and carbon dioxide reduction progressively occur at deeper depths in the profile (Stumm and Morgan 1996). The depth range of any particular reduction reaction may vary from millimeters to meters, depending on the abundance and quality of oxidants and reductants (Kirk 2004), and depth distributions of reduction reactions frequently overlap. Lateral variability in redox conditions from sub-micron to centimeter scales may also be due to soil physical structure and the presence of microsites, as described above.

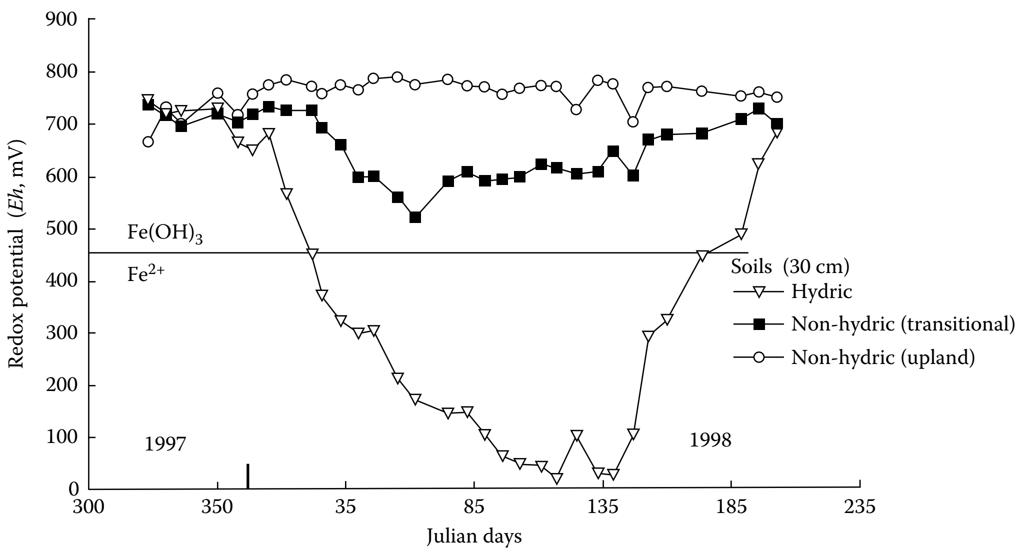


FIGURE 4.3
Comparison of mean redox potential values among three soils (30 cm depth): a hydric soil, a non-hydric soil in the transition to the upland, and in an upland, non-hydric soil. The hydric soil is the only one where the redox potential fell low enough for Fe reduction to occur. The other two soils either did not saturate or were not saturated for a long enough period for Fe reduction to occur.

Interpreting Redox Potential Changes in Nature

Redox potential measurements are made to evaluate changes in soil chemistry. Because of the problems created by the mixed potentials and reaction kinetics discussed earlier, it is safest to base the interpretations of redox data on one or two elements that are abundant in soils and react quickly to changes in redox potential. We will use Fe as the element for interpreting changes in redox potential over time, and focus on the reduction of $Fe(OH)_3$. The first step is to identify the redox potential at which $Fe(OH)_3$ reduces to Fe^{2+}. This redox potential is obtained from the *Eh*/pH diagram shown in Figure 4.1 by using the average pH of the soil measured over time. For the soil shown in Figure 4.2, the average pH was found to be 5.0. From the *Eh*/pH diagram, it can be seen that at this pH $Fe(OH)_3$ reduces to Fe^{2+} when the *Eh* is below 467 mV.

The phase change for $Fe(OH)_3$ to Fe^{2+} is shown in Figure 4.2 by the horizontal line at an *Eh* of 467 mV. The data in Figure 4.2 can be interpreted by considering when and for how long Fe^{2+} was in solution. It can be seen that during most of 1998, Fe^{2+} would have been expected to be in solution. We know from our earlier discussion that if Fe^{2+} is present, we can assume that most dissolved O_2 has been reduced to H_2O, that most NO_3^- present has been denitrified, and that most Mn oxides have been reduced to Mn^{2+}. Microsite reduction and reaction kinetics affect the validity of these assumptions as discussed previously. Phase lines for SO_4^{2-} and CO_2 could also be added to interpret whether these materials were reduced as well. Such interpretations are simple and straightforward, and can be verified by analyzing soil samples with dyes that react with Fe^{2+} or by analyzing water samples for Fe^{2+}.

pH Changes in Reduced Soils

Oxidation–reduction reactions in the anaerobic soil can cause changes in the soil's pH. As shown in Table 4.1, the reducing reactions consume protons, and a change in pH should be expected as a result. Ponnamperuma (1972) showed that the amount of change varies among soils, but, in general, reduction causes the soil pH to shift *toward* 7 but not to necessarily *reach* 7. Reduction in acid soils generally increases the pH, while in alkaline soils it can reduce pH. The amount of pH change can be as high as three pH units following several weeks of submergence although changes of <2 pH units are probably more typical. The degree of change depends on the amount of reduction taking place and is determined by the amount of oxidizable organic tissue, as well as the amount of reducible electron acceptors. According to Ponnamperuma (1972), pH values remain <6.5 in acid soils containing low amounts of organic matter and reducible Fe oxides or hydroxides. In alkaline soils, pH tends to decrease toward 7, possibly due to the production of CO_2. Most acid, organic soils are low in Fe, and submergence would not cause large increases in pH.

Elemental Redox Cycles, Transformations, and Plant Nutrient Pools

Natural wetland systems maintain a wider range of redox reactions than upland ecosystems and transform carbon and plant nutrients among solid, solute, and gaseous forms. As a result, they are capable of recycling key plant elements among the soil, water, and atmosphere. The pools, transformations, and fluxes of organic and inorganic C and the major plant nutrients N, P, and S in freshwater wetlands will be reviewed in this section. Freshwater wetlands will be our focus because they occupy over 90% of the world's wetland area (Giblin and Weider 1992). Representative examples of hydric soils and chemical processes were chosen because it is beyond the scope of this chapter to exhaustively review all wetland systems and possible processing mechanisms for every potential contaminant. More complete reviews can be found in Reddy and DeLaune (2008), Reddy and D'Angelo (1994), Richardson (1999), and Giblin and Wieder (1992) for C, N, P, and S, respectively.

Carbon

Carbon redox transformations represent some of the most important biogeochemical processes within hydric soils because they influence the cycling of a range of elements and regulate environmental quality. These transformations are summarized in the generalized diagram of the carbon cycle for oxidized and reduced zones in a flooded soil shown in Figure 4.4. The diagram is simplified to illustrate the major forms of C and the major processes converting one form into another. Broadly, redox reactions involving carbon in wetlands convert carbon among (a) dissolved and particulate organic matter; (b) inorganic carbon as gaseous carbon dioxide, dissolved carbonate species, and solid carbonate minerals; and (c) methane gas (Figure 4.4a). Wetland soils hold up to one-third of the global soil carbon (Bridgham et al. 2006, Reddy and DeLaune 2008), and C redox transformations within wetlands play important roles for the global carbon cycle.

Within the aerobic zone of wetlands, *photosynthesis* by plants and microbes utilizes atmospheric carbon dioxide and light to form organic matter, reducing C from the 4+ state in CO_2 (Figure 4.4b). Organic matter, in particulate or dissolved forms, is subsequently degraded through *respiration*, coupling oxidation of C to reduction of either O_2 in the aerobic zone (Equations 4.1 through 4.3) or less energetically favorable electron acceptors—such as nitrate, Mn^{4+}, Fe^{3+}, sulfate, and carbon dioxide—in the anaerobic zone (Equations 4.4 through 4.8) (Figure 4.4b). Because microbes gain less energy from anaerobic respiration than aerobic respiration (Table 4.1), organic matter degradation may be slow and it may accumulate within hydric soils (Kirk 2004) depending on soil environmental and biological controls (Schmidt et al. 2011). Globally, the wetland soil C pool is estimated at 513 Pg, but C emissions following human conversion of wetlands to other land uses result in a net loss of 68 Tg C yr^{-1} to the atmosphere (Bridgham et al. 2006).

One important example of anaerobic respiration within wetland soils is carbon dioxide reduction, or *methanogenesis*, whereby organic carbon degradation is coupled with reduction of C in CO_2 to produce CH_4, or methane gas with C in the 4– valence state (Equation 4.8; Figure 4.4b). Methanogenesis occurs under only very reducing conditions (Table 4.1) by a respiring microbial population called methanogens. The derived methane can in turn be oxidized under both aerobic and anaerobic conditions by methanotrophic bacteria. *Methane oxidation* converts methane into carbon dioxide (Figure 4.4b), coupling the oxidation of carbon with reduction of sulfate, nitrate, or other electron acceptors. The production and destruction of methane within wetlands is important for the global carbon cycle because methane is a potent greenhouse gas, and methane emitted from wetlands is a factor influencing global climate change, as discussed below. Methane fluxes from wetlands are modulated by a complexity of physical, chemical, and biological factors (Keller 2011), but on balance, wetlands are net sources of methane to the atmosphere. Recent estimates indicate wetlands emit 105 Tg CH_4 yr^{-1} globally, with the majority coming from freshwater mineral soil wetlands, followed by non-permafrost peatlands and permafrost peatlands (Bridgham et al. 2006).

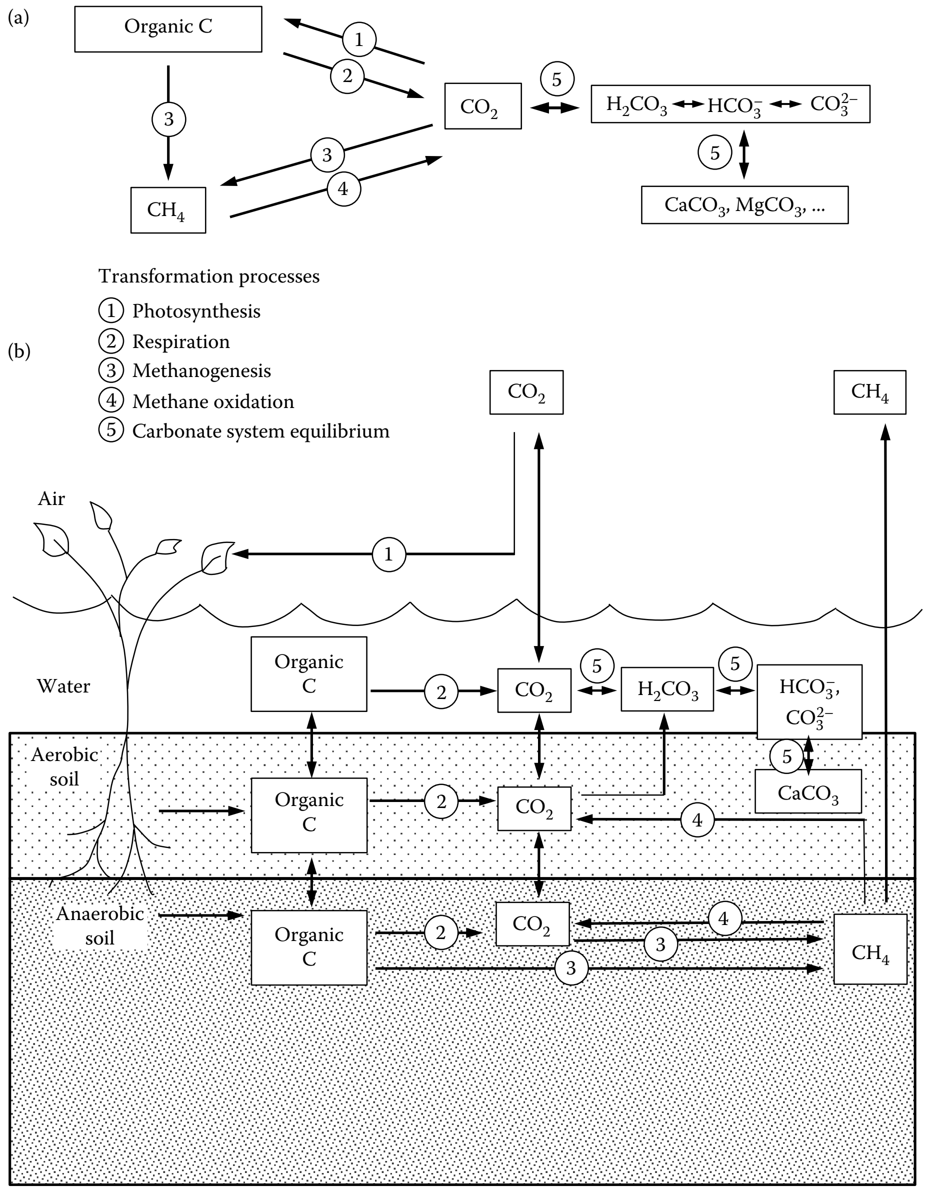


FIGURE 4.4
Schematic of the carbon cycle. The major transformations are shown in (a) while the portions of the soil, as well as water and air in which the transformations occur, are shown in (b). (After Mitsch, W. J. and J. G. Gosselink. 2007. *Wetlands*. 4th ed. John Wiley & Sons, New York, NY.) The C cycle contains gaseous C phases, which allow C to be removed from the wetland.

Carbon dioxide may exchange from the atmosphere into wetland waters and diffuse through the soil system, or it may be produced through respiration and methane oxidation. Carbon dioxide concentrations within hydric soil systems are then governed by *carbonate system speciation and phase changes*, which partition oxidized carbon (4+) among the gas (CO_2), dissolved (H_2CO_3, HCO_3^-, and CO_3^{2-}), and solid (carbonate minerals such as $CaCO_3$ and

$MgCO_3$) phases (Figure 4.4b). Specific partitioning is influenced by pH and the solubility of different carbonate minerals, as well as the respective equilibrium constants of the reactions. However, at circumneutral pH, HCO_3^- is the primary dissolved species, and carbonate minerals are relatively stable (Stumm and Morgan 1996). Conversion of CO_2 to organic matter through photosynthesis and stabilization of carbon within carbonate minerals are important processes for transforming and sequestering atmospheric CO_2 within hydric soils.

Carbon redox transformations within hydric soils regulate environmental quality at both local and global scales. Wetlands play a major role in the sequestration and emissions of potent greenhouse gases, generally serving as a sink for CO_2 but globally delivering up to 40% of the CH_4 to the atmosphere (Reddy and DeLaune 2008, Keller 2011). Organic matter also provides the main source of electrons to drive reduction of electron acceptors, resulting in transformations of pollutants, such as nutrients (e.g., nitrate, nitrite) and trace elements (e.g., arsenic), that may deteriorate surface water or groundwater quality. For instance, organic matter degradation coupled to iron and arsenic reduction within hydric soils has been implicated as one cause of arsenic contamination of well water for 100 million people in Southern Asia (Polizzotto et al. 2008, Neumann et al. 2010). The impacts of wetland redox processes on nutrient cycling are described below.

Nitrogen

Nitrogen transformations in hydric soils are a complex assortment of interrelated processes controlled by microbial activity and the redox status of the soil (Gambrell and Patrick 1978, Reddy and Patrick 1984). These transformations are summarized in the generalized diagram of the nitrogen cycle for oxidized and reduced zones in a flooded soil shown in Figure 4.5. The diagram is simplified, and not all intermediary forms of N are shown. We can assume that the nitrogen cycle contains five basic transformations that can occur in hydric soils (Figure 4.5a). The major N pools or forms in which N occurs in natural freshwater wetlands are: (a) total organic N, which consists of N in plants, microbes, and sediment, (b) available inorganic N in water and sediment (primarily NO_3^- and NH_4^+), and (c) N_2 and N_2O gases. The organic N pool is the largest, with 100–1000 g N m^{-2} occurring primarily as sediment-N (Howard-Williams and Downes 1993). Sediment-N accounts for 80%–90% of the entire N in wetlands. The total plant N pool is roughly an order of magnitude less than total sediment-N. Inorganic-sediment N is another order of magnitude less than the plant pool. Major N inputs into wetlands come from atmospheric deposition (NO_3^- and NH_4^+), N_2 fixation, and groundwater input of NO_3^-.

As shown in Figure 4.5b, *N-Fixation* is the conversion of N_2 gas into NH_3, which is then incorporated into organic tissues. N-fixation rates in wetlands vary from 0.02 to 90 g N m^{-2} yr^{-1} around the world, being lowest in the arctic areas and highest in areas that receive excessive nutrient inputs (Howard-Williams and Downes 1993). N-fixation supplies the majority of N to some wetlands but adds <5% of the N input to others (DeLaune and Patrick 1990, Howard-Williams and Downes 1993). N-fixation in wetlands is accomplished by both blue-green algae and bacteria. Blue-green algae fix nitrogen through photosynthesis and function primarily in water above the soil surface. Other bacteria can fix nitrogen anaerobically and are most abundant in the upper 5 cm of anaerobic soil. Although diffusion of N_2 through water is slow, hydrophytic plants can transport N_2 to their roots as they transport O_2 (Patrick 1982). For this reason, N-fixation rates in flooded soil are greatest in the zone of highest root growth.

Ammonification or mineralization is the conversion of organic-N into ammonium (NH_4^+) by microorganisms. It occurs when organic tissues are oxidized, and the N content of the

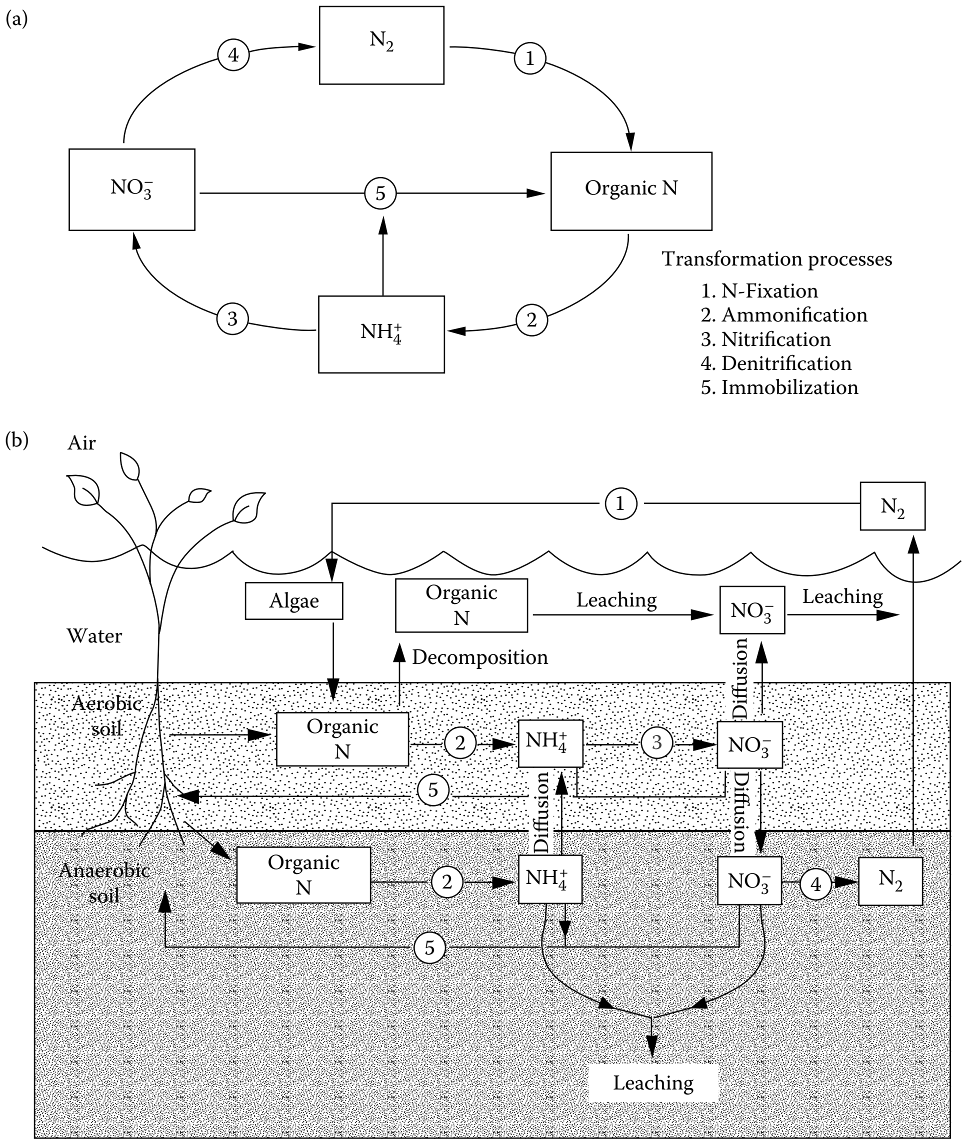


FIGURE 4.5
Schematic of the nitrogen cycle. The major transformations are shown in (a) while the portions of the soil, as well as water and air in which the transformations occur, are shown in (b). The N cycle contains a gaseous N phase, which allows N to be removed from the wetland.

tissue exceeds the requirements of the microbes. Ammonification can occur aerobically or anaerobically (Figure 4.5b), but aerobic ammonification is much faster. Patrick (1982) reported that corn stalks, which are high in N compounds, lost 20% of their weight in 17 days when they were decomposed anaerobically, but lost 37% in the same period when decomposed aerobically. Patrick (1982) also reported that rye grass lost 7% of its weight in 84 days under anaerobic decomposition but lost 17% when decomposed aerobically for 66 days. While anaerobic decomposition is slower, it results in approximately five times the N being maintained as NH_4^+ than occurs under aerobic conditions, in part because nitrification is prevented (Patrick 1982). According to Ponnamperuma (1972) most of the mineralizable N can be converted into ammonium within 2 weeks of submergence in soils with neutral pH, adequate levels of P, and a temperature that is not limiting microbial growth.

Nitrification is the production of NO_3^- from NH_4^+. It occurs in aerobic soils in a two-step process. The ammonium is first converted into nitrite (NO_2^-), and then the nitrite is converted into nitrate (NO_3^-). Both steps in the process are completed by a restricted group of bacteria that include *Nitrosomonas*, *Nitrosococcus*, and *Nitrobacter* species. Nitrification occurs in the aerobic soil zone of hydric soils and around aerated roots growing in anaerobic soil. Rates for nitrification are variable and depend on the quantities of both NH_4^+ and O_2 in the soil. Rates have been reported to range from 0.01 to 0.16 g N m^{-2} day^{-1} (Reddy and D'Angelo 1994). In hydric soils, nitrification rates are limited primarily by the supply of O_2 (Reddy and D'Angelo 1994).

Denitrification is the reduction of NO_3^- to either N_2 or N_2O gases, which is accomplished by a large number of different bacteria (Ponnamperuma 1972). Denitrification is the major mechanism that returns N, originally fixed from the atmosphere, back to the atmosphere. It is a reducing reaction with a rate dependent on the availability of decomposable organic tissues, as well as a supply of NO_3^-. Because NO_3^- is used as an alternate electron acceptor to O_2, denitrification occurs in anaerobic zones of soils or sediments. The anaerobic zones can be the small microsites described earlier. Denitrification rates range from 0.003 to 1.02 g N m^{-2} day^{-1} (Reddy and D'Angelo 1994). *Immobilization* is the conversion of mineral forms of N into plant tissue. Immobilization and ammonification generally occur simultaneously.

Most flooded soils have a thin (<5 cm), oxidized layer at the surface where oxidation reactions occur (Faulkner and Richardson 1989, Moore and Reddy 1994). This layer forms by O_2 diffusing from the atmosphere into the overlying floodwater, and then into the soil (Figure 4.5b) (Howeler and Bouldin 1971). Reduction processes dominate in the anaerobic zone below this oxidized layer. In both reduced and oxidized layers, organic N may be mineralized to NH_4^+. In the reduced layer, NH_4^+ is stable and may be adsorbed to sediment exchange sites or used by both plants and microbes. The thin, oxidized layer in flooded soils is important in N transformations because NH_4^+ may be oxidized to NO_3^- by chemautotrophic bacteria (nitrification) in this layer. Depletion of NH_4^+ in the upper, oxidized layer causes NH_4^+ to diffuse upward in response to the concentration gradient. This diffusion process may be effective from 4 to 12 cm deep.

Nitrate is unstable in reduced zones and is quickly depleted via assimilative reduction (taken up by organisms and used in their tissue) or denitrification, and leaching may further remove nitrate from the system (Figure 4.5b). Redox potential, pH, moisture content, labile (readily oxidizable) C source, and temperature control the rate of NO_3^- reduction. For example, at pHs <6 the reduction of N_2O to N_2 is strongly inhibited. The sequential processes of mineralization, nitrification, and denitrification dominate wetland N cycling and potentially process 20–80 g N m^{-2} yr^{-1} (Bowden 1987).

Phosphorus

The soil P cycle is fundamentally different from the N cycle in that it has no substantial gaseous phase to facilitate the removal of P from wetlands. The lack of the gaseous phase means that P entering a wetland is either stored in wetland sediment and plant tissue or it is in solution and may be carried out of the wetland by flowing water. Phosphorus in soils has a constant valence of 5+, and it is not directly affected by redox processes because it does not change valence. The solubility of P in soils and water, however, is affected by redox processes that dissolve or decompose the compounds that have bonded to P.

A simplified P cycle is shown in Figure 4.6a. Soil P exists in three major forms: organic P, fixed mineral P, and orthophosphate (ortho P). Orthophosphate exists as an anion in the forms of $H_2PO_4^{1-}$, HPO_4^{2-}, and PO_4^{3-}, at pHs of 2–7, 8–12, and >13, respectively. Fixed mineral P consists of ortho-P bound to an oxide or hydroxide containing Al or Fe^{3+}, or bound

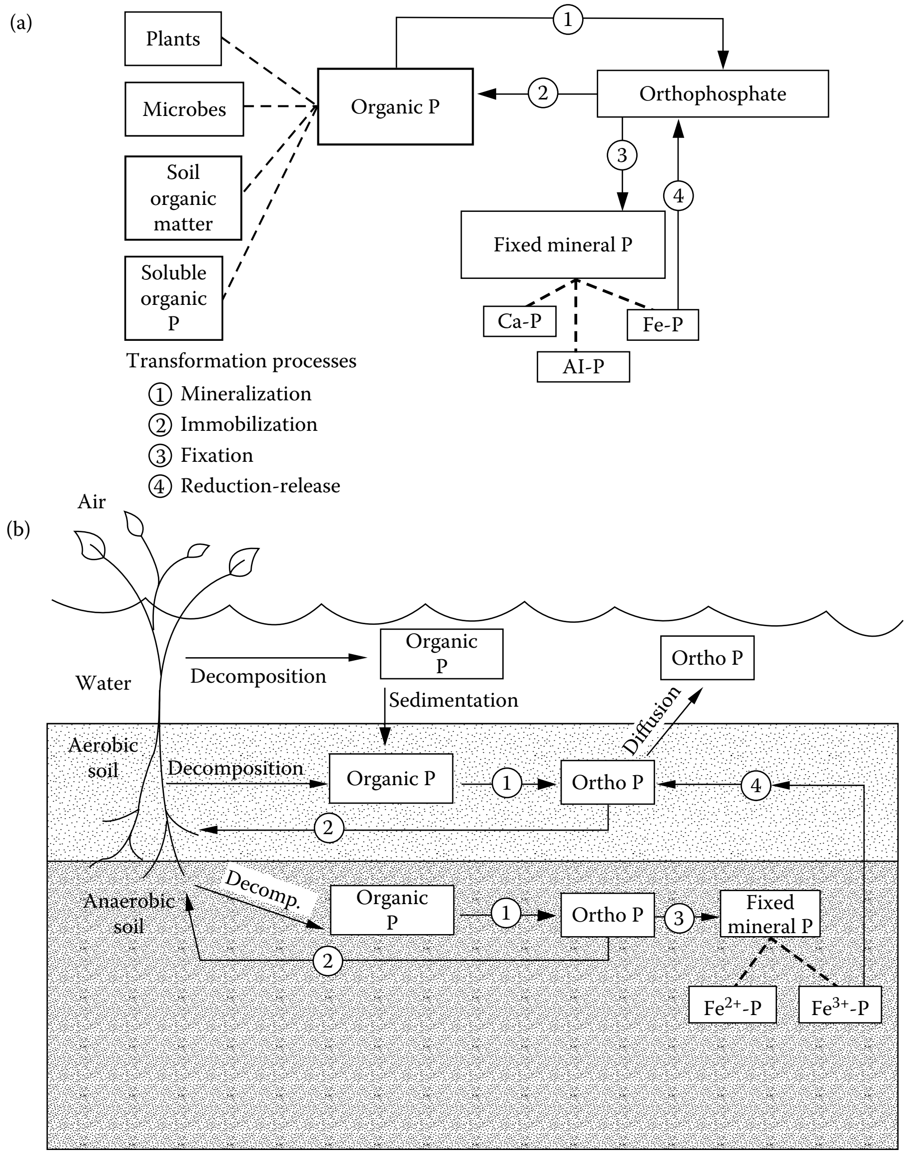


FIGURE 4.6
Schematic of the phosphorus cycle. The major transformations are shown in (a) while the portions of the soil, as well as water in which the transformations occur, are shown in (b). The P cycle does not have a gaseous phase for P, and this causes P to either build up in the wetland soil or to be carried out of the wetland with moving water.

to cations Ca or Mg, all of which are insoluble at certain pHs. Organic forms of P found in plants occur in compounds such as inositol phosphates, phospholipids, and nucleic acids. Insoluble organic P compounds can also be found in sediment in the form of partially decomposed plant tissue. Organic forms of P also include orthophosphate anions that are electrostatically bound to an organic compound such as humic or fulvic acid. These acids are negatively charged, as is the ortho-P, and a bonding of the two is accomplished by a cation bridge that can be Ca, Mg, Al, or Fe. Soluble P may consist of ortho-P anions or as certain organic P forms.

Organic-P and fixed mineral-P comprise approximately 80%–90% of the P in a wetland (Figure 4.6b). Living plants store most of the remaining P, leaving very little in the water column (Richardson 1999). Whether P in the soil is in the organic form or fixed mineral form depends largely on the type of soil present. Most soil P (>95%) in organic soils is in the organic form with cycling among P-forms controlled by biological forces (i.e., microbes and plants). In mineral soils, most of the soil P may be bound to minerals containing Al, Fe, Ca, or Mg.

The transformation of fixed mineral P into soluble orthophosphate is controlled by the interaction of redox potential and pH (Holford and Patrick 1979, Sah and Mikkelsen 1986). In acid soils (pHs 4–7) ortho-P is preferentially adsorbed onto Fe and Al oxides and hydroxides. At pHs <4, the Al and Fe oxide and hydroxides dissolve, and any ortho-P bonded to them can be released to the soil solution (Lindsay 1979). Likewise, Fe^{3+} phosphates such as strengite ($FePO_4 \cdot 2H_2O$) can also dissolve if the redox potential falls low enough to reduce the ferric Fe to Fe^{2+} (Moore and Reddy 1994). Aluminum does not change valence in soils and is unaffected by redox potential. Precipitation as insoluble Ca- or Mg-phosphate minerals or adsorption to carbonates is the dominant transformations at pHs >7. These are the predominant forms of mineral P in arid soils. Just as Al, Ca, and Mg do not participate in redox reactions, so the solubility of Ca- or Mg-P complexes is determined by soil or water pH rather than *Eh* (Moore and Reddy 1994).

The amount of P that a soil can adsorb is directly related to the amount of oxalate-extractable (amorphous) Al and Fe^{3+} oxides and hydroxides (Richardson 1985, Reddy and D'Angelo 1994). The P sorption capacity of an oxidized soil may increase during flooding and reduction due to formation of amorphous ferrous hydroxides, which have a greater surface area and more sorption sites than the more crystalline, oxidized, ferric forms (Holford and Patrick 1979). While organic materials can also absorb P, the amount of organic material in the soil is generally not as good a predictor of P adsorbing capacity as the amounts of amorphous Fe and Al oxides and hydroxides. Because soils have a finite amount of Fe and Al oxides and hydroxides, their P sorbing capacity is limited. Once it is exceeded, no more P can be retained in the wetland, and it is then kept in solution and carried out of the wetland with water.

In addition to the amount of P a wetland can hold, the rate at which P is added to a wetland also determines whether P is retained within or exported from a wetland. Richardson (1999) presented an analysis of 125 wetland sites and showed that the approximate maximum rate at which a wetland can absorb P is <1 g P m^{-2} yr^{-1}. When this rate of P input is exceeded, wetlands do not absorb P fast enough to remove it from solution, and the excess P is exported from the wetland with flowing water. Sediment accretion processes control the long-term P removal capability of wetland ecosystems. Sediment accretion rates for peats have been estimated to store <1 g P m^{-2} yr^{-1} (Richardson 1999).

There is little direct uptake of phosphate from the water column by emergent wetland vegetation because the soil is the major source of nutrients (Richardson 1999). Growing vegetation is a temporary nutrient-storage compartment resulting in seasonal exports following plant death. The long-term role of emergent vegetation is to transform inorganic P to organic forms. Microorganisms play a definite role in P cycling in wetlands, but the microbial pool is small in terms of P storage (10%–20% of total P) (Richardson and Marshall 1986).

Sulfur

The S cycle has been studied less than those for N and P in freshwater wetlands. The major forms of S found in these wetlands are shown in Figure 4.7a. Most (>70%) soil S occurs in

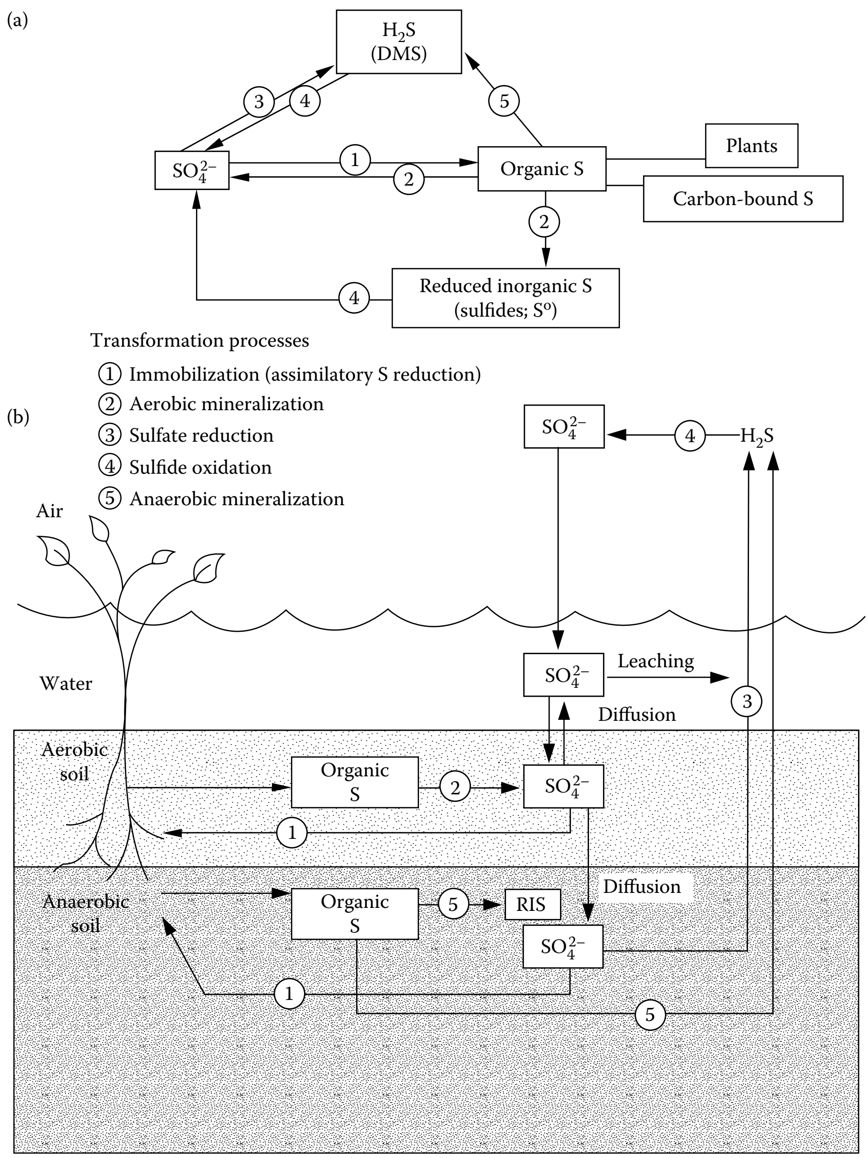


FIGURE 4.7
Schematic of the sulfur cycle. The major transformations are shown in (a) while the portions of the soil, as well as water and air in which the transformations occur, are shown in (b). The sulfur cycle includes a gaseous phase as does the nitrogen cycle, allowing S to be removed from the wetland.

an organic form (Wieder and Lang 1986, 1988). The remaining inorganic S is distributed as *reduced inorganic sulfides* (RIS) such as pyrite (<10%), *dissolved* SO_4^{2-} (<20%), and *gases* consisting of H_2S and dimethyl sulfide (<5%) (Giblin and Weider, 1992).

Sulfur transformations are biologically mediated and, like P and N transformations, are affected by the interaction between redox potential and pH. In aerated portions of the hydric soil (Figure 4.7b) immobilization, oxidation of inorganic sulfide and elemental sulfur, and mineralization of organic S to inorganic SO_4^{2-} are the dominant processes (Giblin and Weider 1992). Under reducing conditions, sulfate reduction transforms SO_4^{2-} to H_2S during respiration by obligate anaerobic bacteria. The H_2S formed by sulfate reduction can be released to the atmosphere or can react with organic matter, providing another pathway for converting inorganic-S into organic-S. With SO_4^{2-} reduction and sufficient Fe^{2+}, iron sulfides (FeS and FeS_2) can form; pyrite (FeS_2) formation requires alternating (either temporally or spatially) anaerobiosis with limited aeration.

Despite the small size of the inorganic pool, this fraction is the most important for S cycling, retention, and mobility. Fluxes through the inorganic pool dominate S cycling in wetlands that have high SO_4^{2-}, inputs such as those usually associated with wastewater additions. Wieder and Lang (1988) calculated that 3.5–4 times as much inorganic S was processed compared to the organic S pool through alternating SO_4^{2-} reduction and sulfide/sulfur oxidation. This has important implications for wetland S cycles because S inputs are primarily SO_4^{2-} from atmospheric deposition and either natural or amended hydrologic sources. Sulfate retention by aerobic, mineral soils is dominated by the same adsorption mechanisms involved in PO_4^{3-} retention. However, adsorbed SO_4^{2-} is displaced by PO_4^{3-} on the exchange sites, but PO_4^{3-} is not displaced by SO_4^{2-}.

Significant fluxes of S to the atmosphere cause the wetland to function as a transformer as opposed to a true sink. Studies reviewed by Giblin and Weider (1992) show that H_2S emission from freshwater wetlands is generally <200 mg S m^{-2} yr^{-1}, and losses from dimethyl sulfide emission are approximately the same. Sulfate reduction rates in saltwater marshes can be 10 times higher than those in freshwater wetlands due to a greater amount of SO_4^{-2} input (Giblin and Weider 1992).

Summary

Hydric soils differ from upland soils in that they are anaerobic in their upper 30 cm for some period during most years. The anaerobic conditions develop when oxidation–reduction reactions occur in the soil that transfer electrons from donor to acceptor atoms. These reactions require that a soil: (1) be saturated with slowly moving or stagnant water, (2) have oxidizable organic C, and (3) have an active microbial population. The soils must be saturated to exclude atmospheric oxygen gas, which is a strong electron acceptor. Organic C is needed to supply the electrons used in the reduction process. An active microbial population transfers the electrons from donors to acceptors as they respire and oxidize organic tissues.

The major electron acceptors include the elements O, N, Mn, Fe, S, and C. The elements are reduced in this order based on thermodynamic constraints. The presence of oxygen in the form of O_2 gas can keep all other elements from being reduced. The reduction of these elements can be monitored in soils by measuring the oxidation–reduction (redox) potential, using dye solutions, or with IRIS tubes. Redox potential is an electrical measurement, where the voltage developed between two electrodes can be related to the chemistry of the soil solution. Dyes such as α,α′-dipyridyl react with reduced forms of Fe and allow its detection in the field. IRIS tubes are PVC tubes coated with a Fe oxyhydroxide paint that dissolves in reduced soils.

Oxidation–reduction reactions affect soil color by causing organic C to accumulate and Fe or Mn oxides to become concentrated or depleted in portions of the soil. The hydric soil field indicators that are used to identify hydric soils are all formed by oxidation–reduction reactions. These reactions are also responsible for the cycling of N, P, and S in soils by transforming these elements into organic and inorganic forms. This cycling has important implications for maintaining the quality of fresh waters and impacting the buildup of greenhouse gases in the atmosphere.

References

Baas-Becking, L. G. M., I. R. Kaplan, and D. Moore. 1960. Limits of the natural environment in terms of pH and oxidation–reduction potentials. *J. Geol.* 68: 243–284.

Beauchamp, E., J. T. Trevors, and J. W. Paul. 1989. Carbon sources for bacterial denitrification. *Adv. Soil Sci.* 10: 113–142.

Bohn, H. L., B. L. McNeal, and G. A. O'Connor. 1985. *Soil Chemistry.* John Wiley & Sons, New York, NY.

Bouma, J. 1983. Hydrology and soil genesis of soils with aquic moisture regimes. In L. P. Wilding, N. E. Smeck, and G. F. Hall (Eds.) *Pedogenesis and Soil Taxonomy. I. Concepts and Interactions.* Elsevier Science Publishers B.V., Amsterdam, The Netherlands, pp. 253–281.

Bowden, W. B. 1987. The biogeochemistry of nitrogen in freshwater wetlands. *Biogeochemistry* 4: 313–348.

Bridgham, S. D., J. P. Megonigal, J. K. Keller, N. B. Bliss, and C. Trettin. 2006. The carbon balance of North American wetlands. *Wetlands* 26(4): 889–916.

Castenson, K. L. and M. C. Rabenhorst. 2006. Indicator of reduction in soil (IRIS): Evaluation of a new approach for assessing reduced conditions in soil. *Soil Sci. Soc. Am. J.* 70: 1222–1226.

Childs, C. W. 1981. Field test for ferrous iron and ferricorganic complexes (on exchange sites or in watersoluble forms) in soils. *Aust. J. Soil Res.* 19: 175–180.

Clark, M. H. and C. L. Ping. 1997. Hydrology, morphology, and redox potentials in four soils of South Central Alaska. In M. J. Vepraskas and S. W. Sprecher (Eds.) *Aquic Conditions and Hydric Soils: The Problem Soils*. SSSA Spec. Publ. No. 50, Soil Sci. Soc. Am., Madison, WI, pp. 113–131.

Collins, J. F. and S. W. Buol. 1970. Effects of fluctuations in the *Eh*–pH environment of iron and/or manganese equilibria. *Soil Sci.* 110: 111–118.

Couto, W., C. Sanzonowicz, and A. De O. Bacellos. 1985. Factors affecting oxidation–reduction processes in an Oxisol with a seasonal water table. *Soil Sci. Soc. Am. J.* 49: 1245–1248.

Crozier, C. R., I. Devai, and R. D. DeLaune. 1995. Methane and reduced sulfur gas production by fresh and dried wetland soils. *Soil Sci. Soc. Am. J.* 59: 277–284.

DeLaune, R. D. and W. H. Patrick, Jr. 1990. Nitrogen cycling in Louisiana Gulf Coast brackish marshes. *Hydrobiologia* 199: 73–79.

Faulkner, S. P. and C. J. Richardson. 1989. Physical and chemical characteristics of freshwater wetland soils. In D.A. Hammer (Ed.) *Constructed Wetlands for Wastewater Treatment*. Lewis Publishers, Chelsea, MI, pp. 41–72.

Faulkner, S. P., W. H. Patrick, Jr., and R. P. Gambrell. 1989. Field techniques for measuring wetland soil parameters. *Soil Sci. Soc. Am. J.* 53: 883–890.

Firestone, M. K. and E. A. Davidson. 1989. Microbiological basis of NO and N_2O production and consumption in soil. In M. O. Andreae and D. S. Schimel (Eds.) *Exchange of Trace Gases between Terrestrial Ecosystems and the Atmosphere.* John Wiley & Sons Ltd., Chichester, U.K.

Gambrell, R. P. and W. H. Patrick, Jr. 1978. Chemical and microbiological properties of anaerobic soils and sediments. In D. D. Hook and R. M. M. Crawford (Eds.) *Plant Life in Anaerobic Environments.* Ann Arbor Science Publishers, Inc., Ann Arbor, Michigan, pp. 375–423.

Germida, J. J. 1998. Transformations of sulfur. In D. M. Sylvia, J. J. Furhmann, P. G. Hartel, and D. A. Zuberer (Eds.) *Principles and Applications of Soil Microbiology.* Prentice-Hall, Upper Saddle River, NJ, pp. 346–368.

Giblin, A. E. and R. K. Wieder. 1992. Sulphur cycling in marine and freshwater wetlands. In R. W. Howarth, J. W. B. Stewart, and M. V. Ivanov (Eds.) *Sulphur Cycling on the Continents. Wetlands, Terrestrial Ecosystems, and Associated Ecosystems.* SCOPE 48. John Wiley & Sons, New York, NY, pp. 85–117.

Gilman, K. 1994. *Hydrology and Wetland Conservation.* John Wiley & Sons, New York, NY.

Heaney, S. I. and W. Davison. 1977. The determination of ferrous iron in natural waters with 2,2′-bipyridyl. *Limnol. Oceanogr.* 22: 753–760.

Holford, I. C. R. and W. H. Patrick, Jr. 1979. Effects of reduction and pH changes on phosphate sorption and mobility in an acid soil. *Soil Sci. Soc. Am. J.* 43: 292–297.

Howard-Williams, C. and M. T. Downes. 1993. Nitrogen cycling in wetlands. In T. P. Burt, A. L. Heathwaite, and S. T. Trudgill (Eds.) *Nitrate: Processes, Patterns, and Management.* John Wiley & Sons, New York, NY, pp. 141–167.

Howeler, R. H. and D. R. Bouldin. 1971. The diffusion and consumption of oxygen in submerged soils. *Soil Sci. Soc. Am. Proc.* 35: 202–208.

Jenkinson, B. J. and D. P. Frnazmeier. 2006. Development and evaluation of Fe coated tubes that indicate reduction in soils. *Soil Sci. Soc. Am. J.* 70: 183–191.

Keller, J. K. 2011. Wetlands and the global carbon cycle: What might the simulated past tell us about the future? *N. Phytol.* 192: 789–792.

Kirk, G. 2004. *The Biogeochemistry of Submerged Soils.* John Wiley & Sons, New York, NY.

Lindsay, A. L. 1979. *Chemical Equilibria in Soils.* John Wiley & Sons, New York, NY.

McBride, M. B. 1994. *Environmental Chemistry of Soils.* Oxford Univ. Press, New York, NY.

Meek, B. D., A. J. McKenzie, and L. B. Gross. 1968. Effects of organic matter, flooding time, and temperature on the dissolution of iron and manganese from soil *in situ*. *Soil Sci. Am. Proc.* 32: 634–638.

Megonigal, J. P., S. P. Faulkner, and W. H. Patrick, Jr. 1996. The microbial activity season in southeastern hydric soils. *Soil Sci. Soc. Am. J.* 60: 1263–1266.

Mitsch, W. J. and J. G. Gosselink. 2007. *Wetlands.* 4th ed. John Wiley & Sons, New York, NY.

Moore, P. A. and K. R. Reddy. 1994. Role of *Eh* and pH on phosphorus geochemistry in sediments of Lake Okeechobee, Florida. *J. Envir. Qual.* 23: 955–964.

Neumann, R., K. Ashfaque, A. Badruzzaman, M. Ali, J. Shoemaker, and C. Harvey. 2010. Anthropogenic influences on groundwater arsenic concentrations in Bangladesh. *Nat. Geosci.* 3: 46–52.

Parkin, T. B. 1987. Soil microsites as a source of denitrification variability. *Soil Sci. Soc. Am. J.* 51: 1194–1199.

Patrick, W. H. 1982. Nitrogen transformations in submerged soils. In F. J. Stevenson (Ed.) *Nitrogen in Agricultural Soils.* Agron. Monogr. 22. ASA, CSSA, and SSSA, Madison, WI.

Patrick, W. H., Jr., R. P. Gambrell, and S. P. Faulkner. 1996. Redox measurements of soils. In D. L. Sparks (Ed.) *Methods of Soil Analysis: Part 3—Chemical Methods.* 3rd ed. Soil Sci. Soc. Am., Madison, WI, pp. 1255–1276.

Polizzotto, M. L., B. D. Kocar, S.G. Benner, M. Sampson, and S. Fendorf. 2008. Near-surface wetland sediments as a source of arsenic release to ground water in Asia. *Nature* 454: 505–508.

Ponnamperuma, F. N. 1972. The chemistry of submerged soils. *Adv. Agron.* 24: 29–96.

Rabenhorst, M. C. 2008. Protocol for using and interpreting IRIS tubes. *Soil Survey Horiz.* 49: 74–77.

Rabenhorst, M. C. 2012. Simple and reliable approach for quantifying IRIS tube data. *Soil Sci. Soc. Am. J.* 76: 307–308.

Rabenhorst, M. C., W. D. Hively, and B. R. James. 2007. Measurements of soil redox potential. *Soil Sci. Soc. Am. J.* 73: 668–674.

Reddy, K. R. and R. D. DeLaune. 2008. *Biogeochemistry of Wetlands: Science and Applications*. CRC Press, Boca Raton, FL.

Reddy, K. R. and E. M. D'Angelo. 1994. Soil processes regulating water quality in wetlands. In W. J. Mitsch (Ed.) *Global Wetlands. Old World and New.* Elsevier Science Publishers, B.V., Amsterdam, The Netherlands, pp. 309–324.

Reddy, K. R. and W. H. Patrick, Jr. 1984. Nitrogen transformations and loss in flooded soils and sediments. *CRC Crit. Rev. Environ. Control* 13: 273–309.

Richardson, C. J. 1985. Mechanisms controlling phosphorus retention capacity in freshwater wetland. *Science* 228: 1424–1427.

Richardson, C. J. 1999. The role of wetlands in storage, release, and cycling of phosphorus on the landscape: A 25-year retrospective. In K. R. Reddy, G. A. O'Connor, and C. L. Schelske (Eds.) *Phosphorus Biogeochemistry in Subtropical Ecosystems.* Lewis Publ., Boca Raton, FL, pp. 47–68.

Richardson, C. J. and P. E. Marshall. 1986. Processes controlling movement, storage, and export of phosphorus in a fen peatland. *Ecol. Monogr.* 56: 279–302.

Richardson, J. L. and F. D. Hole. 1979. Mottling and iron distribution in a Glossoboralf Haplaquoll hydrosequence on a glacial moraine in northwestern Wisconsin. *Soil Sci. Soc. Am. J.* 43: 552–558.

Sah, R. N. and D. S. Mikkelsen. 1986. Transformations of inorganic phosphorus during the flooding and draining cycles of soil. *Soil Sci. Soc. Am. J.* 50: 62–67.

Schmidt, M. W. I., M. S. Torn, S. Abiven, T. Dittmar, G. Guggenberger, I. A. Janssens, M. Kleber. et al. 2011. Persistence of soil organic matter as an ecosystem property. *Nature* 478: 49–56.

Soil Survey Staff. 1999. *Soil Taxonomy: A Basic System of Soil Classification for Making and Interpreting Soil Surveys.* 2nd ed. USDA, Natural Resources Conservation Service, Agr. Handbook No. 436. U.S. Govt. Printing Office, Washington, DC.

Sparks, D. L. 2003. *Environmental Soil Chemistry.* 2nd ed. Academic Press, San Diego, CA.

Stumm, W. and J. J. Morgan. 1996. *Aquatic Chemistry: Chemical Equilibria and Rates in Natural Waters.* 3rd ed. John Wiley & Sons, New York, NY.

Theis, T. L. and P. C. Singer. 1973. The stabilization of ferrous iron by organic compounds in natural waters. In P. C. Singer (Ed.) *Trace Metals and Metal–Organic Interactions in Natural Waters.* Ann Arbor Science Publ., Ann Arbor, MI, pp. 178–192.

Turner, F. T. and W. H. Patrick, Jr. 1968. Chemical changes in waterlogged soils as a result of oxygen depletion. *Trans. 9th Int. Cong. Soil Sci.* 4: 53–65.

Wieder, R. K. and G. E. Lang. 1986. Fe, Al, Mn, and S chemistry of sphagnum peat in four peatlands with different metal and sulfur input. *Water Air Soil Pollut.* 29: 309–320.

Wieder, R. K. and G. E. Lang. 1988. Cycling of inorganic and organic sulfur in peat from big run bog, West Virginia. *Biogeochemistry* 5: 221–242.

Zhi-guang, L. 1985. Oxidation-reduction potential. In Y. Tian-red (Ed.) *Physical Chemistry of Paddy Soils*. Springer-Verlag, Berlin, pp. 1–26.

5

Biology of Wetland Soils

Anne E. Altor

CONTENTS

Introduction

The development of wetland soil characteristics is mediated by organisms that inhabit the soil ecosystem. These organisms, in turn, are influenced by the physiochemical environment, especially climate, hydrology, and water chemistry. Wetland nutrient status, determined in part by hydrogeomorphic setting, helps determine the productivity and composition of the plant community and the soil microbial processes that characterize the system (Balasooriya et al. 2008). The biota that inhabit wetland soils drive the formation of hydric soil characteristics, nutrient cycling, primary and secondary production, decomposition and accumulation of organic matter, habitat quality, and biodiversity in these ecosystems (Holguin et al. 2001; Batzer et al. 2006; Van der Valk 2006).

Wetland soils are often described as biologically "stressful" environments because they are characterized by periodic to continuous anaerobic conditions and may be subjected to salinity and scouring (Naidoo et al. 1992; Blom 1999). The organisms that inhabit wetland soils have evolved anatomical and physiological adaptations to these stressors (Otte 2001). As in other ecosystems, the soils of wetlands are inhabited by microorganisms (bacteria, archaea, and fungi), plants, invertebrates, and vertebrates. This chapter will examine biologically mediated processes in freshwater and estuarine wetland soils, including nutrient cycling and organic matter dynamics; adaptations of plants and invertebrates to anaerobic and saline soil environments; and interactions among organisms and processes, including ecosystem engineering, in wetland soil ecosystems.

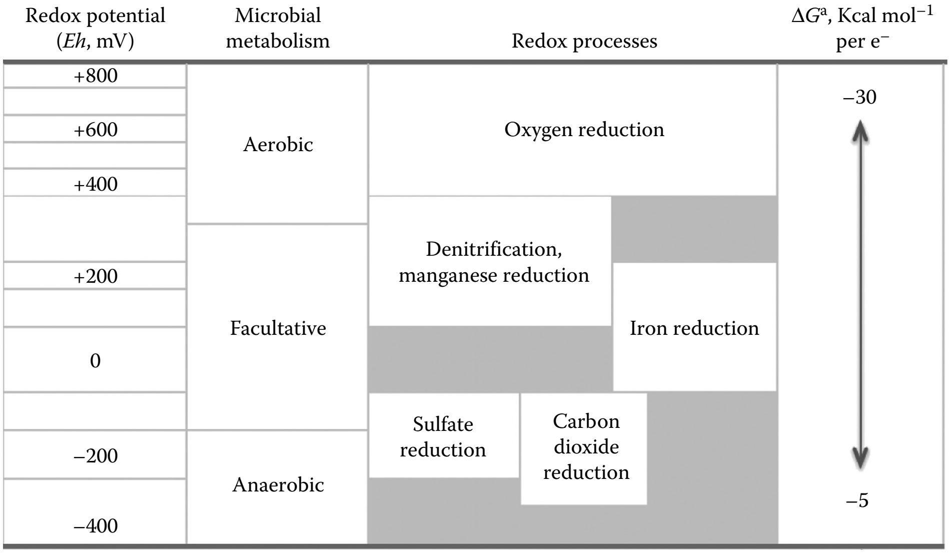
Microorganisms

Wetland soil microbes include bacteria, archaea, and fungi. Microorganisms in soils and pore water drive the cycling of essential nutrients including carbon, nitrogen, phosphorus, iron, and sulfur (Ponnamperuma 1972; Paerl and Pinckney 1996; Gutknecht et al. 2006). Bacterial and archaeal biomass contributes to secondary production in surface sediments (Moran and Hodson 1992; Buesing and Gessner 2006). Although fungi are predominantly aerobic organisms, they are important decomposers of aboveground and submerged wetland plant detritus (Gulis et al. 2006; Gessner et al. 2007) and often exist in symbiotic mycorrhizal associations with wetland plant roots (Thormann et al. 1999; Turner et al. 2000).

Various decomposition and nutrient cycling processes occur in close proximity in wetland soils along gradients in oxygen availability, redox potential, pH, and types and concentrations of electron donors and acceptors (Chapter 4) (Paerl and Pinckney 1996). When soils are flooded, aerobic microbial metabolism quickly depletes the pool of available oxygen; oxygen is replenished slowly under saturated conditions, where its diffusion is approximately 10^4 times slower than in air. Facultative and anaerobic microorganisms use oxidized molecules including NO_3^-, SO_4^{2-}, and Fe^{3+} as electron acceptors along a gradient of thermodynamic potential and energy yield (Figure 5.1). Lower oxidation–reduction (redox) potential (reduced conditions) indicates decreased availability of oxidized molecules, lower potential for redox reactions to occur, and lower energy yield from the reactions (Reddy et al. 2000).

Bacteria and Archaea

Bacteria and archaea comprise a diverse assemblage of organisms and functional groups. Bacteria can form multicellular functional colonies (Shapiro 1988), while archaea are known to exist only as unicellular organisms (Kletzin 2007). Both taxa form consortia in which the metabolic processes of individual species or strains complement one another and modify the microenvironment in ways that facilitate microbial community metabolism (e.g., nitrification and denitrification; Paerl and Pinckney 1996). Because of the difficulty in isolating or culturing distinct species, bacteria and archaea are often described by their functional roles or are discussed at the generic taxonomic level. The abundance of bacteria and archaea



[a]Gibbs free energy, assuming the reduction reaction is coupled to the oxidation reaction:
$1/4\ CH_2O + 1/4\ H_2O \rightarrow 1/4\ CO_2 + H^+ + e^-$, and $\Delta G = -RT\ \text{In}(K)$.

FIGURE 5.1
Generalized schematic of microbial metabolism in relation to oxidation-reduction potential, substrates used in redox processes, and approximate energy yield for the various reactions. (Adapted from Schlesinger, W. H. 1997. *Biogeochemistry: An Analysis of Global Change*, 2nd ed. Academic Press, San Diego, CA; Reddy, K. R., E. M. D'Angelo, and W. G. Harris. 2000. *Handbook of Soil Science*. CRC Press, Boca Raton, FL, pp. G89–G119; Sigg, L. 2000. *Redox: Fundamentals, Processes and Applications*. Springer, Berlin, Germany.)

is high in wetland ecosystems. Counts of 10^9–10^{10} bacteria/g of dry soil were recorded in a seasonal freshwater marsh in Florida, USA (Ipsilantis and Sylvia 2007), comparable to estimates made from other freshwater and estuarine wetland sediments (Rublee 1982; Moran et al. 1987). Wetland bacteria include groups that obtain nutrients by transforming organic compounds (chemoheterotrophic, e.g., denitrifying and iron-reducing bacteria) or inorganic molecules (chemolithotrophic, e.g., nitrifying and iron-oxidizing bacteria) (Batzer and Sharitz 2006; Weber et al. 2006b). Phototrophic bacteria (e.g., cyanobacteria) are found in the water column and on sediments (Reddy and DeLaune 2008).

Wetland soils are dominated by facultative or obligate anaerobic organisms, although aerobic microbes are active in oxygenated soil zones (e.g., the rhizosphere) and play important roles in resupplying oxidized substrates, such as Fe^{3+} and NO_3^-, for respiration (Gutknecht et al. 2006). The high energy return of aerobic oxidation (38 mol ATP/mol glucose) enables single microbial taxa to mineralize organic matter completely to carbon dioxide; energy yield is lower under anaerobic conditions (as low as 2 mol ATP/mol glucose), and syntrophic microbial consortia, in which taxa specialize on specific substrates, carry out the mineralization processes (Atlas and Bartha 1998; Megonigal et al. 2004). Facultative anaerobes reduce substrates including Fe^{3+} and NO_3^- in reactions that yield slightly less energy than aerobic respiration (Boon 2006). The reduction of molecules including CO_2 and SO_4^{2-} by obligate anaerobes yields approximately 7- to 17-fold less energy than aerobic respiration, depending on the electron source (Roden and Jin 2011). Anaerobic decomposition processes include hydrolysis of complex organic compounds to simple molecules (e.g., sugars, amino acids) by extracellular enzymes; fermentation reactions, in which organic molecules serve

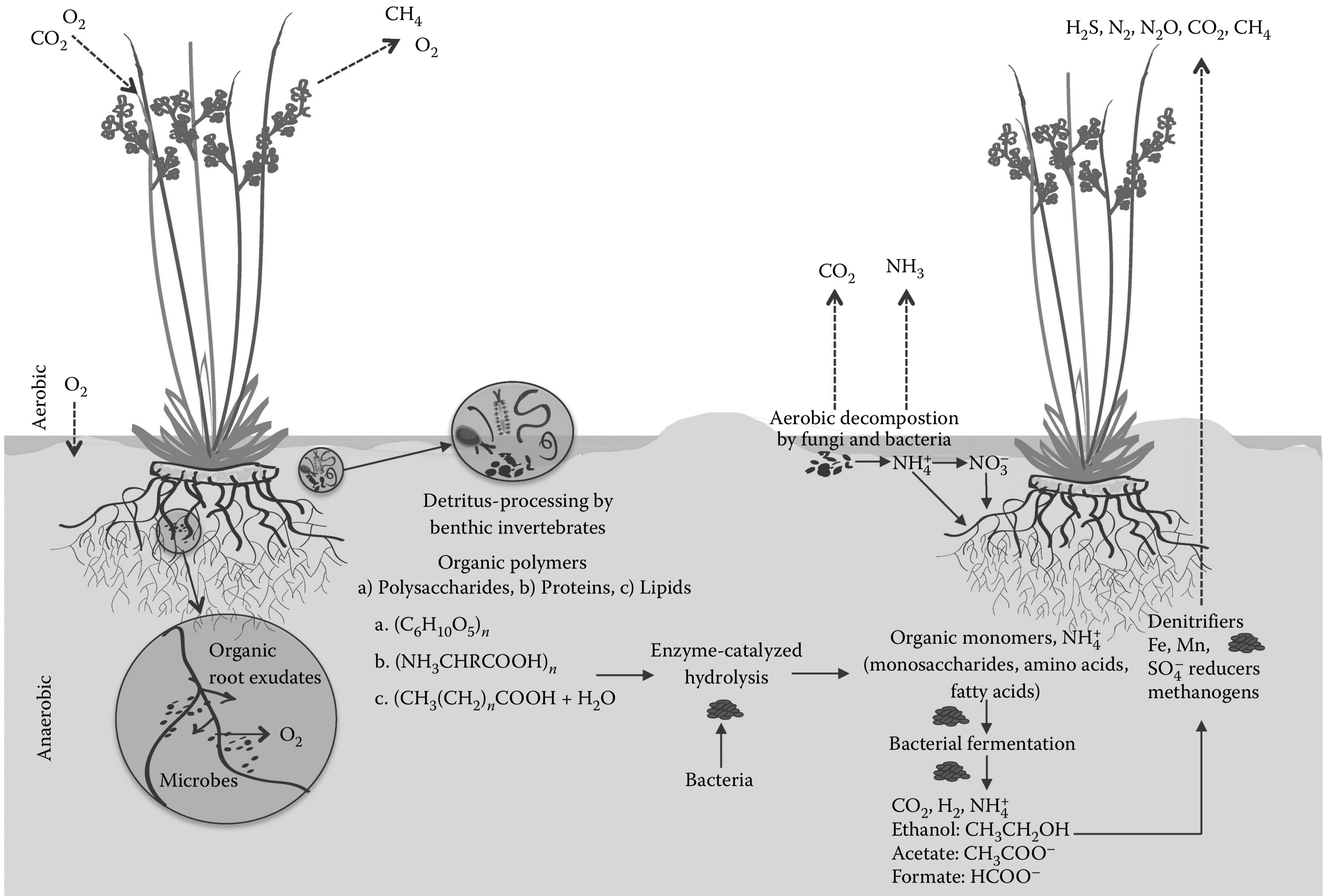


FIGURE 5.2
Decomposition of organic matter in a wetland. Typical invertebrates and microbial reaction pathways are depicted. Dotted lines indicate exchange of gases between plants, soil, and atmosphere.

as both electron donors and acceptors; and mineralization of organic fermentation products via reduction of substrates such as Fe^{3+}, SO_4^{2-}, or CO_2 (Figure 5.2) (Lovley and Phillips 1986; Weston and Joye 2005). Fermentation produces low-molecular-weight dissolved organic molecules that can be mineralized by microbial consortia to terminal end products including Fe^{2+}, N_2, H_2S, NH_4^+, and CH_4 (Weston et al. 2006).

Archaea are best known for their ability to inhabit "extreme" environments such as hydrothermal vents, but molecular techniques have revealed their broad distribution across ecosystems including wetlands (Chaban et al. 2006). Wetland soils support a diversity of archaea, of which methanogens may be the most abundant (Pazinato et al. 2010). Populations of 1.1–8.3 × 10^9 methanogens/g soil were recorded in Chinese marshes and peatlands (Liu et al. 2011). Other archaea include methane-oxidizing (Raghoebarsing et al. 2006), iron-oxidizing (Weber et al. 2006a), ammonia-oxidizing (Hofferle et al. 2010), and nitrifying (Zhu et al. 2011) groups.

Fungi

Fungi are heterotrophic organisms that are ubiquitous in nature. Although the majority of fungi grow best in the presence of oxygen, many facultative and some obligately anaerobic strains with diverse metabolic strategies have been identified (Tabak and Cooke 1968; Wainwright 1988). Facultative and anaerobic fungi are found in both freshwater and saltwater wetland sediments (Tonouchi 2009; Mohamed and Martiny 2011), and some facultative soil fungi use denitrification and dissimilatory nitrate reduction to ammonium (DNRA) to obtain nutrients and energy (Shoun and Tanimoto 1991; Zhou et al. 2002). Mycorrhizae, mutualistic associations between fungi and plant roots, are found in the majority of upland plant species (Amaranthus 1998; Bever et al. 2001; Langley and Hungate 2003). A growing body of research demonstrates that mycorrhizae are also common in freshwater (Thormann et al. 1999; Bauer et al. 2003; Wolfe et al. 2006) and oligohaline to saline wetland ecosystems (Carvalho et al. 2001; Sengupta and Chaudhuri 2002).

Free-Living Fungi

Fungi are best known for their role as decomposers, especially of wood and other plant litter (Wainwright 1992). Wetland fungi are most abundant on standing dead plant material and can comprise a significant proportion of the microbial biomass in wetlands (Gessner et al. 2007). Within the litter layer, fungal activity is greatest in oxygenated zones and declines with depth and redox potential (Padgett and Celio 1990; Mansfield and Barlocher 1993). Fungi are important decomposers of exposed and submerged wetland plant litter and may condition litter for colonization by bacteria, which dominate decomposition in sediments (Newell et al. 1995).

Mycorrhizal Fungi

Mycorrhizal fungi are mutualistic species that form associations with plant roots and include organisms from all fungal phyla (Harley 1989). In a mycorrhizal relationship, plant roots provide fungi with carbon, and the fungi facilitate nutrient exchange between plant roots and the soil (Siddiqui and Pichtel 2008). Networks of fungal filaments (hyphae) extend from plant roots into the soil matrix where they access a larger volume of resources than would be accessible to uncolonized roots (Parniske 2008). Hyphae have a strong affinity for phosphorus and incorporate this nutrient into polyphosphates within vacuoles;

polyphosphates are translocated to hyphae in plant roots where a series of reactions make phosphorus available to the plant (Siddiqui and Pichtel 2008). Fungal hyphae can also take up amino acids, ammonium, and nitrate (Parniske 2008; Hobbie et al. 2009). Mycorrhizal fungi absorb simple sugars inside plant roots; they convert these sugars into lipids for storage or for export to extraradical hyphae where they serve as substrates for respiration and growth (Siddiqui and Pichtel 2008).

Oxygen supplied by plant roots may enable mycorrhizal fungi to survive in anaerobic wetland sediments (Miller and Bever 1999). In an experimental study, mycorrhizae-colonized lowland ecotypes of black gum (*Nyssa sylvatica*) seedlings under submerged conditions, and mycorrhizal colonization was positively correlated with above- and belowground *N. sylvatica* biomass (Keeley 1980). Research in salt marshes has shown that mycorrhizal responses to soil nutrients can influence the competitive ability and zonation of vegetation. A field experiment by Daleo et al. (2008) revealed that aboveground growth of the mycorrhizal plant *Spartina densiflora* exceeded that of non-mycorrhizal *Spartina alterniflora* at ambient nutrient levels but that plant growth dynamics were reversed when nutrients were added. The competitive advantage obtained by mycorrhizal plants may be negated or reversed under nutrient enrichment, if non-mycorrhizal plants released from nutrient limitation produce aboveground biomass at a greater rate than plants that allocate carbon to root growth or mycorrhizal fungi (Daleo et al. 2008). The presence of mycorrhizae in wetland plants has been widely confirmed. However, the ecological and functional roles of these symbioses have only begun to be clarified (Gutknecht et al. 2006; Weishampel and Bedford 2006).

Microorganisms and Nutrient Cycling

Carbon

Although they comprise a relatively small (±6%) proportion of the Earth's surface (Mitsch and Gosselink 2000; Whigham 2009), wetlands contain a significant proportion of carbon sequestered in the biosphere. Current analyses estimate that wetlands contain up to one-third of the carbon sequestered in soils globally, with the highest concentrations found in peatlands (Dise 2009, Chapter 6). Microbial activity decreases with decreasing temperature, which leads to slow rates of decomposition in cold climates where many peatlands form. Acidic conditions in many peatland soils also suppress bacterial growth, for which the optimal pH range is approximately 6–8 (Atlas and Bartha 1998). Microorganisms process carbon between organic and inorganic forms, thus performing a key role in the global carbon cycle. Aerobic decomposition occurs in unsaturated soils and in oxygenated microsites around plant roots and invertebrate burrows when soils are inundated (Kristensen et al. 2000). However, anaerobic processes are responsible for the majority of decomposition in inundated and saturated soils (Reddy et al. 2000). Slow rates of decomposition under anaerobic conditions lead to greater accumulation of soil organic matter in wetlands compared with many terrestrial ecosystems (Gorham 1991; Craft and Richardson 1993; Chmura et al. 2003). Methanogenesis (anaerobic production of CH_4) occurs in wetland soils and has been widely studied because of methane's importance as a greenhouse gas (see Laanbroek 2010 and Bridgham et al. 2013 for recent reviews).

Decomposition occurs in multiple stages in wetland soils (Figure 5.2). A thin oxygenated layer is often present at the surface of submerged soils, where the diffusion of O_2 from the water column resupplies oxygen depleted by aerobic respiration. Fungi and aerobic or facultative bacteria colonize the plant material in oxygenated sediments and secrete extracellular enzymes that catalyze hydrolysis of organic compounds (Su et al. 2007).

Invertebrate shredders such as crabs can contribute to decomposition by breaking up the plant material, which increases the surface area and accessibility of litter to detritivores (Middleton and McKee 2001; Su et al. 2007). Scrapers such as mayfly larvae consume benthic algae and detritus and excrete waste products that are used by other trophic groups; filtering and gathering collectors, including copepods and many dipteran larvae, process fine particulate organic matter that is subsequently utilized by microorganisms (Merritt et al. 1996). Organic matter varies in quality, degradability, and accessibility to microorganisms. Labile compounds (e.g., amino acids and sugars) are easily decomposed, while more complex or high-molecular-weight compounds (e.g., cellulose, lignin, humic acids) degrade slowly and accumulate over time as refractory organic matter (Brinson et al. 1981).

Nitrogen

The largest pool of nitrogen in wetlands is generally contained in the soil (Bowden 1987), and the majority of nitrogen in wetland soils is incorporated into organic molecules including amino acids, proteins, and recalcitrant (humic) compounds (White and Reddy 2009). Nitrogen is biologically cycled through organic and inorganic forms and through solid, dissolved, and gas phases (Figure 5.3). Plant roots take up NH_4^+, NO_3^-, and organic N monomers, and vascular tissue of the plants can transport N_2O and N_2 gases from the soil to the atmosphere (Schimel and Bennett 2004; Kirk and Kronzucker 2005; White and Reddy 2009). Microbial communities use diverse metabolic pathways to process nitrogen for energy and growth.

The primary transformations of nitrogen in wetland soils include mineralization (ammonification), nitrification, denitrification, DNRA, and anaerobic ammonium oxidation (ANAMMOX; Chapter 4). Nitrogen fixation, carried out by free-living or symbiotic heterotrophic and photosynthetic bacteria, can be important in wetlands with limited external inputs of nitrogen, such as some peatlands (Waughman and Bellamy 1980; Howarth et al. 1988; Vymazal 2007). As organic matter is decomposed, nitrogen is released as NH_4^+, which can be taken up by plants or oxidized to NO_3^- in aerobic soil microsites by nitrifying bacteria (Reddy et al. 2000). NH_4^+ can also be oxidized under anaerobic conditions to N_2 (Burgin and Hamilton 2007). Nitrifying bacteria include the genera *Pseudomonas*, *Nitrosomonas*, *Nitrobacter*, and *Alcaligenes* (Castignetti and Hollocher 1982). Nitrification and denitrification are closely coupled in wetland soils as a result of microscale variations in redox status and along O_2 gradients near soil-water boundaries, plant root zones, and bioturbated areas (Seitzinger et al. 2006; White and Reddy 2009).

Denitrification, the microbial reduction of NO_3^- to N_2 or N_2O gases, is a primary pathway for the removal of N from wetlands (Clément et al. 2002; White and Reddy 2009). Another process, DNRA, transforms NO_3^- to less-mobile, biologically available NH_4^+ (Burgin and Hamilton 2007). Aerobic and facultative anaerobic denitrifying bacteria include species from the genera *Pseudomonas*, *Aeromonas*, *Vibrio*, *Acinetobacter*, and *Alcaligenes* (Prakasam 1984; Herbert and Nedwell 1990). Ambient rates of denitrification ranged from 0.2 to 10.9 mg N/(m^2 h) in salt marshes and riparian wetlands, and from 0.3 to 17.0 mg N/(m^2 h) in created and constructed wetlands receiving agricultural drainage or municipal wastewater (Hernandez and Mitsch 2007). DNRA may be favored over denitrification in reduced soils that are rich in labile carbon, because many bacteria that carry out DNRA (e.g., *Clostridium*) are obligate anaerobes, and the energy yield from DNRA is greater than that of denitrification (Kelso et al. 1999; Burgin and Hamilton 2007). Other bacteria that carry out DNRA include facultative anaerobic (e.g., *Citrobacter*, *Enterobacter*) and aerobic (e.g., *Pseudomonas*) genera (Rutting et al. 2011). Nitrate is used as an electron acceptor

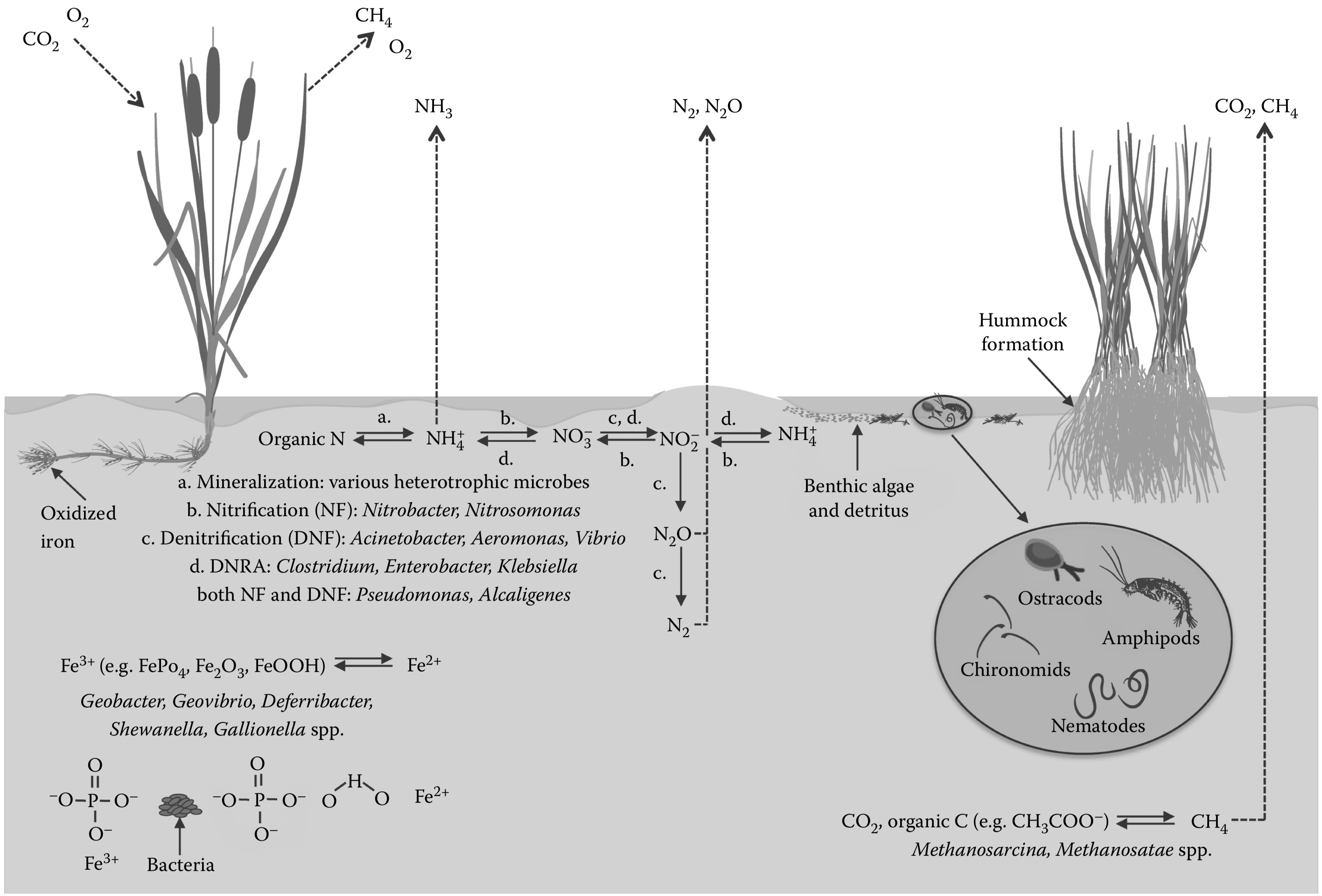


FIGURE 5.3
Freshwater marsh soil ecosystem. Typical soil invertebrates and nutrient transformations are illustrated (intermediate steps in reactions not shown). Dotted lines indicate exchange of gases between plants, soil, and atmosphere.

for cellular metabolism rather than for cell synthesis in both denitrification and DNRA (Tiedje 1988). ANAMMOX is a microbial process in which reduction of NO_3^- or NO_2^- is coupled to oxidation of NH_4^+ (Vymazal 2007). Other substrates and products have been documented from ANAMMOX reactions, including the reduction of Fe^{3+} to Fe^{2+} with oxidation of NH_4^+ to NO_2^- (Clément et al. 2005). There is wide variability in the proportion of N_2 produced by ANAMMOX, from <1% in tidal creeks and salt marshes (Koop-Jakobsen and Giblin 2010) to >80% in brackish coastal sediments (Burgin and Hamilton 2007).

Phosphorus

In wetland soils, phosphorus is found mainly as free orthophosphates (PO_4^{3-}), as exchangeable and soluble mineral-bound phosphates, in crystallized mineral forms such as $Fe_3(PO_4)_2$, as polyphosphates, and as a component of organic matter (Figures 5.3 and 5.4) (Ponnamperuma 1972; Reddy et al. 1999; Sundareshwar et al. 2001). Soil texture and composition, pH, hydrology and redox status, and biological activity are the major factors that affect phosphorus dynamics in soils (Richardson and Vaithiyanathan 2009). A substantial proportion of wetland soil phosphorus can be contained in the microbial biomass (Wright et al. 2001), although recalcitrant organic compounds form the major sink (Reddy and D'Angelo 1994). Mineralization of organic matter releases phosphate, which is then available for biological uptake or abiotic immobilization (Reddy and D'Angelo 1994). Reducing conditions can facilitate the release of phosphorus from ferric phosphates or from microbial biomass as aerobic organisms die or facultative organisms hydrolyze stored polyphosphates to produce ATP (Davelaar 1993; Reddy et al. 1999). The available phosphorus also increases when wetland soils dry out and microbial biomass and other organic matter is oxidized (Boon 2006). In estuarine wetlands, sulfide binds reduced iron (Fe^{2+}), which limits its diffusion and oxidation to Fe^{3+}, which in turn reduces the quantity of substrates available to adsorb phosphate (Caraco et al. 1990). The greater availability of PO_4^{3-} observed in estuarine, compared with freshwater, sediments might be explained in part by the decreased availability of Fe^{3+} in sulfate-rich sediments (Jordan et al. 2008).

Iron and Manganese

Iron is a significant constituent of mineral soils, averaging approximately 4% by mass (Fageria et al. 2002), and biochemical transformations of iron form the primary redoximorphic features that are utilized as indicators of hydric soils (Vepraskas 1992). Reduction of oxidized iron (Fe^{3+}) and manganese (Mn^{4+}) are energetically favorable metabolic pathways under anaerobic conditions after nitrate has been reduced. Reddy and DeLaune (2008) describe three groups of microorganisms involved in iron (and manganese) transformations in wetland soils: Fe- or Mn-reducing bacteria (e.g., *Geobacter* spp.) that transfer electrons to Fe^{3+} and Mn^{4+} from organic or inorganic molecules; Fe- or Mn-oxidizing bacteria (e.g., *Gallionella* spp.) that transfer electrons from Fe^{2+} and Mn^{3+} to O_2, SO_4^{2-}, or nitrogen oxides (Figures 5.3 and 5.4); and bacteria (e.g., *Leptothrix* spp.) that precipitate Fe and Mn intra- and extracellularly. The charged surfaces of bacterial cells induce binding and precipitation of iron; these "biomineralization" processes provide an oxidant source (Fe^{3+}) that can accept electrons from mineralization of organic matter under anaerobic conditions (Konhauser 1998). Substantial amounts of organic carbon can be oxidized via Fe^{3+} reduction in wetland sediments. For example, reduction of Fe^{3+} in sediments of a beaver-created wetland (Alabama, USA) accounted for as much mineralization of

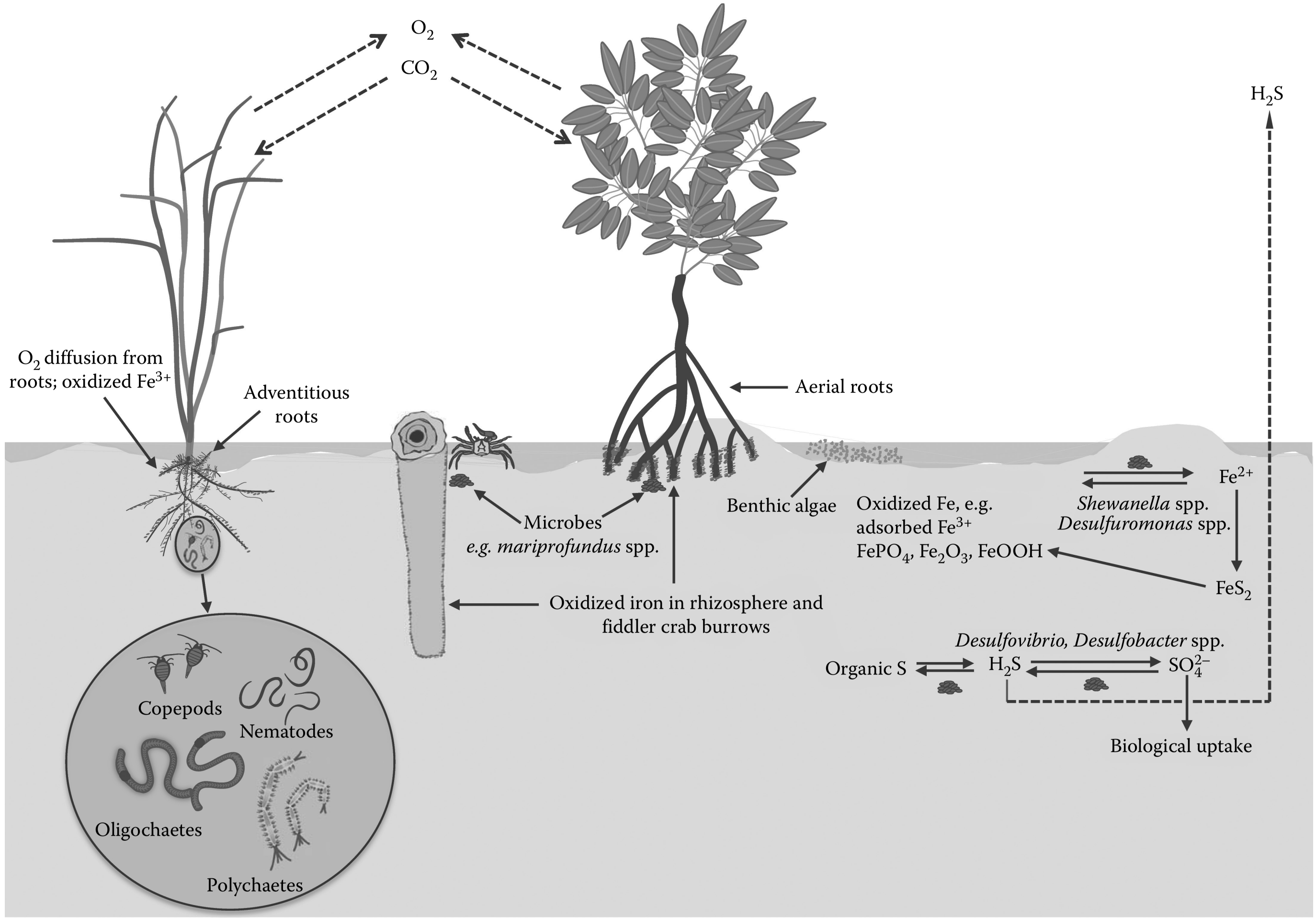


FIGURE 5.4
Estuarine wetland (e.g. salt marsh or mangrove) soil ecosystem. Typical soil invertebrates and nutrient transformations are illustrated (intermediate steps in reactions not shown). Dotted lines indicate exchange of gases between plants, soil, and atmosphere.

organic carbon as methanogenesis (Roden and Wetzel 1996). Microbial Fe^{3+} reduction was an important pathway for carbon oxidation in bioturbated areas of a salt marsh, where reactive Fe^{3+} was regenerated by fiddler crab activity (Hyun et al. 2007).

Sulfur

Sulfur is an important component of ecosystem metabolism and nutrient cycling in salt marshes, mangrove forests, and other estuarine wetlands (Figure 5.4). The average concentration of sulfate in seawater is approximately 2.65 g/kg, while the average concentration in rivers is much less (0.01 g/kg) (Mackenzi and Garrels 1966). Measurements of sulfate reduction, CO_2 production, and O_2 uptake in salt marsh soil cores from Great Sippewissett Marsh, Massachusetts, USA, indicated that approximately 50% of decomposition was coupled to sulfate reduction (Howes et al. 1984). In another study, >90% of carbon mineralization was mediated by sulfate reduction in sediments of a salt marsh in Georgia, USA (Kostka et al. 2002). Sulfate reduction occurred to a greater depth in sediments that were bioturbated by fiddler crabs than in sediments without crabs (Kostka et al. 2002).

Plant roots help drive sulfur cycling in estuarine wetlands. Bacterial abundance and sulfate reduction rates are higher around plant roots than in bulk sediments (Kristensen and Alongi 2006), presumably because of the availability of labile organic compounds released from roots. The carbon-rich, semi-oxygenated conditions in the root zone facilitate rapid oxidation of Fe^{2+} and subsequent SO_4^{2-} reduction and precipitation of FeS_2 (Kristensen and Alongi 2006). In some salt marsh soils, rates of sulfate reduction correspond positively to the density of vegetation (Koretsky et al. 2003).

Vegetation

As in other ecosystems, plant communities are distinguishing characteristics of wetlands, and vegetation influences soil structure and composition. The addition of organic matter from wetland primary production contributes to hydric soil formation and vertical soil accretion and is the foundation of complex food webs and soil biological processes (Nyman et al. 1993; Batzer et al. 2006; Jones et al. 2009). Plant roots and rhizomes stabilize the soil, enhance soil porosity, and provide habitat and surface area for microorganisms and invertebrates (Brix 1997; Ehrenfeld et al. 2005). Vegetation reduces the velocity of water flow, traps sediments, shades soils, and facilitates soil-building processes (Craft and Casey 2000; Leonard et al. 2002; Ellery et al. 2003).

Plants add organic matter to the soil through senescence, root death, passive diffusion of solutes from living root cells (exudation), metabolic secretion of polymers (mucilage), and delivery of carbon compounds to mycorrhizae (Jones et al. 2009). Above- and belowground plant production is the foundation of the soil organic matter pool (Craft et al. 1988; Chen and Twilley 1999; Dise 2009). Organic compounds secreted by plant roots include simple and complex sugars, amino acids and proteins, organic acids, alcohols, and hormones (Nguyen 2003). Because of these carbon inputs, the rhizosphere generally contains large concentrations of microorganisms and invertebrates (Bonkowski et al. 2009) and is a primary locus of soil biological activity (Crow and Wieder 2005). The release of oxygen from plant roots creates small-scale gradients in redox potential, which enables complementary microbial functions (e.g., nitrification and denitrification) to occur simultaneously (White and Reddy 2009).

Plant Adaptations to Anaerobic Soils

Plant species that thrive in wetlands have evolved morphological and metabolic adaptations to inundated or saturated soils (Table 5.1). In addition to low levels of oxygen, anaerobic microbial processes generate by-products that can be toxic to plants, such as reduced iron and hydrogen sulfide (Snowden and Wheeler 1993; Reddy and DeLaune 2008). Morphological adaptations in wetland vegetation include development of porous tissues (aerenchyma) that allow oxygen to diffuse to the rhizosphere; shallow rooting and production of adventitious roots; buttresses, pneumatophores, hypertrophied lenticels, prop and drop roots; hummock formation that effectively raises the elevation of plant roots; and convective gas flow between the air and soil through plant tissues (Figures 5.3 and 5.4). Metabolic adaptations in wetland plants include increased growth rates, anaerobic metabolic pathways, and increased production of antioxidants that minimize oxidative cell damage (Table 5.1). Passive exchange of soil and atmospheric gases also takes place through broken aboveground stems.

TABLE 5.1

Major Plant Adaptations to Anaerobic Soils

Morphological Adaptation	Function	Suggested References
Adventitious root production	Increases oxygen capture and transport; enhances nutrient uptake	Rich et al. (2012)
Aerenchyma	Allows more rapid gas transfer between shoots and roots or rhizomes, to maintain root respiration and oxygenate rhizosphere	Armstrong et al. (1991), Jackson and Armstrong (1999)
Aerial and lateral root growth	Increases root surface area, enhancing aeration and physical stability	Gill and Tomlinson (1977), Cronk and Fennessy (2001)
Buttresses	Provides physical support or enhances aeration	Varnell (1998), Mendelssohn and Batzer (2006)
Hummock or tussock formation	Creates an effective increase in elevation	Cronk and Fennessy (2001), Lawrence and Zedler (2011)
Pneumatophores (knees)	Increases ventilation of root system	Cronk and Fennessy (2001)
Hypertrophied lenticels	Enhances gas exchange between external atmosphere and plant cells, in stems and woody tissue	Mendelssohn and Batzer (2006)
Pressurized (convective) throughflow	Increases rate of gas exchange between plant shoots and roots, which increases oxygenation and detoxification of rhizosphere	Armstrong et al. (1991), Jackson and Armstrong (1999)
Root dimorphism	Thick, aerenchyma-rich soil roots maintain oxygen supply; thin aquatic roots take up nutrients	Koncalova (1990)
Shallow rooting	Increases surface area of roots in better oxygenated, superficial sediments	Tiner (1999)
Metabolic Adaptation	**Function**	**Suggested References**
Anaerobic metabolic pathways, for example, fermentation	Production of ATP from stored carbohydrates enables maintenance of cell growth and function	Mendelssohn et al. (1981), Pezeshki et al. (1993)
Increased growth rate/stem elongation	Maintains plant–air contact for access to adequate oxygen and carbon dioxide	Ridge (1987), Blom (1999)
Increased production of antioxidants in roots and rhizomes	Detoxifies harmful compounds produced when roots are re-exposed to oxygen.	Crawford and Braendle (1996), Blokhina et al. (2003)

Aerenchyma are gas-filled intercellular spaces that are formed by lysis or splitting of cells, particularly in tissues of the leaf and the cortex (Schussler and Longstreth 1996). Oxygen and other gases are transported between shoots and the root zone through aerenchyma, facilitating aerobic root respiration and oxidation of toxic compounds that form under anaerobic conditions (Jackson and Armstrong 1999). Aerenchyma provide a pathway for gas exchange via convective throughflow, which is a more rapid exchange than molecular diffusion. Convective throughflow, also called pressurized ventilation, occurs when a pressure gradient exists between the root and the shoot tissue that causes gases to circulate between the soil and air (Colmer 2003). Radial oxygen loss is concentrated at the growing tips of adventitious roots in some wetland plants by the production of a dense layer of thickened cells along the rest of the root that prevent O_2 loss; this physical restriction of gas flow between root cells and soil helps maintain oxygen supply within the bulk root while allowing oxygen loss at the meristems, which enables growing root tips to penetrate deeper into anaerobic soil (Visser et al. 2000). To balance the need for nutrient uptake with the need to control oxygen loss, some wetland plants show root dimorphism, having both thick, aerenchyma-packed soil roots that maintain oxygen supplies and finely branched, thin aquatic roots that take up nutrients (Koncalova 1990).

Swollen trunks called buttresses occur in some wetland trees, including cypress (*Taxodium*) and tupelo (*Nyssa*) species, in response to flooding. The shape and extent of butt flare correspond to hydroperiod and maximum high water level over time (Kurz and Demaree 1934; Varnell 1998). The specific functions of buttresses have not been definitively resolved; they may enhance aeration of the stem or improve tree stability. Cypress knees (conical extensions of lateral roots that rise above the water surface) are a type of aerial root (pneumatophore) that have been hypothesized to enhance aeration, provide mechanical support to the plant, or to provide carbohydrate storage. In aquatic mangrove (*Rhizophora*) species, pneumatophores form "arches and columns" above the soil surface and lateral, aerenchyma-filled branches where they contact the soil. The root columns are packed with lenticels, which provide sites for gas exchange with underground roots (Gill and Tomlinson 1977).

Anaerobic conditions stimulate some wetland plants to produce ethylene, which induces shoots and petioles to elongate above the water surface. This adaptation enables plants to maintain uptake of O_2 and CO_2 for root respiration and photosynthesis, respectively (Ridge 1987). Under anaerobic conditions, fermentation of stored carbohydrates in roots and rhizomes provides ATP for cell synthesis and metabolism. Some fermentation products (e.g., ethanol) are converted into toxic compounds (e.g., acetaldehyde) when aerobic conditions are restored (Blokhina et al. 2003). Increased production of antioxidants in roots and rhizomes, and diffusion of ethanol out of rhizomes help minimize toxicity in some species (Crawford and Braendle 1996).

Plant Adaptations to Salinity

Estuarine wetlands are affected by the additional stress of salinity. Salinity poses at least two challenges for plants: osmotic stress resulting from the lower water potential of saltwater and toxicity from high concentrations of Na^+ and Cl^- (Cronk and Fennessy 2001; Jenks and Hasegawa 2005). Salinity in the rhizosphere creates an ionic concentration gradient that favors diffusion of water out of the plant tissue, and uptake of Na^+ in excess over K^+ (a required nutrient) can lead to nutrient deficiency (Botella et al. 2005). In addition, high concentrations of sulfide in estuarine soils can inhibit plant physiological functions, especially production of photosynthetic enzymes and nitrogen uptake (Bradley and Morris

TABLE 5.2

Major Plant Adaptations to Salinity

Adaptation	Function	Suggested References
Compartmentalization of Na^+ and Cl^- in vacuoles	Regulates osmotic potential; reduce toxicity of salt ions	Hasegawa et al. (2000)
Succulence	Increases volume per unit surface area, leading to lower ionic concentration (salt dilution)	Naidoo and Rughunanan (1990), Ogburn and Edwards (2010)
Synthesis of compatible solutes	Protects osmotic balance	Hasegawa et al. (2000), Munns and Tester (2008)
Casparian strips	Creates a barrier to salt uptake in root cells	Hajibagheri et al. (1985), Cronk and Fennessy (2001)
Maintenance of very low water potential	Promotes filtration of salts at root surface as water is drawn into cells	Waisel et al. (1986)
Preferential transport of K^+	Absorbs nutrient K^+ while excluding excess Na^+	Rains (1972)
Salt glands	Secretes toxic ions out of leaves	Fahn (1988), Mendelssohn and Batzer (2006)
Shedding	Conserves energy and removes accumulated salts	Munns and Tester (2008)

1990; Cronk and Fennessy 2001). Plant species that thrive in saline environments have evolved mechanisms to exclude or secrete salt, to maintain favorable water potential, or to conserve water (Table 5.2). Salt exclusion is achieved by physical and metabolic mechanisms. Casparian strips (bands of molecules deposited on cell walls) develop in root tips in some salt-tolerant plant species (Hajibagheri et al. 1985); these thickened cell walls create a barrier to the uptake of salt ions. Low water potential in plant tissues enables a filtration process to occur in which water is drawn into root cells and salts are left behind (Waisel et al. 1986). Transport structures in cell membranes of some salt-tolerant plants have a higher affinity for K^+ than for Na^+ and preferentially transport K^+ into the cell (Rains 1972). Some species can selectively transport ions out of root cells or to vacuoles or older leaves (Levinsh 2006). Succulent plants compartmentalize and dilute salts in enlarged, water-filled vacuoles, thus minimizing toxicity and maintaining favorable osmotic potential (Ogburn and Edwards 2010). Some plants produce or accumulate nontoxic "compatible" solutes such as sugars and charged metabolic by-products, which regulate water movement into and out of cells and reduce the toxicity of salts to enzymes (Hasegawa et al. 2000). Secretion of Na^+ and Cl^- from salt glands (specialized epidermal structures called trichomes) is a mechanism for removing salts to the outside surface of the plant (Fahn 1988). Plants experiencing salt toxicity may shed mature tissues more rapidly than unaffected plants, thus conserving energy and removing accumulated ions (Munns and Tester 2008). Metabolic adaptations that enable plants to inhabit saline environments require cellular energy and represent a tradeoff for survival (Volkmar et al. 1998).

Soil Fauna

Invertebrates

Invertebrates are an important part of heterotrophic food webs in wetlands, and these secondary producers influence soil structure and biogeochemical cycling via bioturbation and organic matter processing (Merritt et al. 1996; Middleton and McKee 2001; Van der Valk 2006). Wetland soils are inhabited by detritivores and predatory, grazing, gathering, and filtering species, some of which are unique to wetlands (Wissinger 1999). The composition of benthic invertebrate communities is determined by hydrology and soil physiochemical characteristics (especially O_2 availability), vegetation, spatial heterogeneity, and connectivity with the surrounding landscape (Wissinger 1999). Major invertebrate phyla that are important in wetland soils include Platyhelminthes (flatworms), Nematoda (unsegmented roundworms), Mollusca (snails), Annelida (segmented worms and leeches), and Arthropoda (insects, crustaceans) (Wissinger 1999; Batzer et al. 2006). Many of the micro- and meio-invertebrate phyla are understudied because of the difficulty of collecting, identifying, and preserving specimens (Thorp et al. 2010).

Platyhelminthes (Flatworms)

The flatworms include micro- (mostly <1 mm) and macroturbellarians (up to approximately 30 mm). Microturbellarians have been reported from the sediments of temporary wetlands (Eitam et al. 2004). Some species have developed desiccation-resistant eggs, and desiccation may be required for release from diapause in these species (Kolasa and Tyler 2010). Flatworms are mainly predatory and form a link between micro- and macrofaunal food webs. Their functional niche as consumers of zooplankton, algae, and other invertebrates is thought to play an important role in benthic community structure by regulating the density of their prey (Kolasa and Tyler 2010).

Nematoda (Unsegmented Roundworms)

Nematodes are a diverse and abundant group of unsegmented roundworms; they include bacterivores, fungivores, plant and animal parasites, and omnivores (Wu et al. 2008). Nematodes are among the most abundant meiofauna in wetland sediments ($2–23 \times 10^6/m^2$) and generally inhabit the upper 4 cm (Wieser and Kanwisher 1961; Poinar 2010). Some roundworms are tolerant of both anaerobic and saline conditions and contribute to the breakdown of detritus in salt marsh sediments (Teal 1962). The estimated biomass of nematode populations in salt marshes in the United Kingdom and the United States varies from 2 to 18 g/m^2 (wet weight basis) (Warwick and Price 1979). In salt marshes, sediment-dwelling nematodes link anaerobic and aerobic components of the food web by grazing on detritus-processing microorganisms and by providing a food source for macroinvertebrates and fish.

Mollusca: Gastropoda

Snails are important macroinvertebrates in intertidal and other mudflat soils and can alter the physical and trophic structure of these ecosystems. Mud snails compete with other invertebrates for microinvertebrate and diatom food sources, and their foraging activities can disrupt the habitat of other taxa such as annelids (Kelaher et al. 2003). Snails segregate according to burrow depth within the upper 10 cm of sediments (Jensen and Kristensen 1990). Some species (e.g., *Littoraria irrorata*) graze on salt marsh vegetation and, in turn, serve as a major food source for animals such as birds and crabs. Experimental exclusion of snail predators from salt marsh plots on Sapelo Island, Georgia, revealed that *L. irrorata*, when not controlled by predators, could decimate vegetation and reduce the marsh to a mudflat within 2 years (Silliman and Bertness 2002).

Annelida (Segmented Worms and Leeches)

Phylum Annelida includes three groups of segmented invertebrates: polychaete and oligochaete worms, and hirudinida (leeches), all of which are found in wetlands. The distribution of annelids in wetland sediments depends on moisture, pH, and salinity, with different species inhabiting niches along the continuum of these variables (Beylich and Graefe 2002). Except for leeches, wetland-dwelling annelids are burrowing organisms that digest organic matter and excrete mineral material while feeding on sediments, detritus, microorganisms, and microscopic benthic algae (Rouse 2001; Gillett et al. 2007). Leeches are top predators in some benthic communities where they feed on chironomids, amphipods, mollusks, and other annelids, and in some cases are ectoparasites of vertebrate animals (Govedich et al. 2010). Polychaetes are abundant in marine and estuarine ecosystems and form a dominant component of the benthic invertebrate community (Rouse 2001). Densities $>2 \times 10^5/m^2$ were reported for intertidal mudflat polychaetes in a California wetland (Levin 1984). Oligochaetes (e.g., tubificids and earthworms) are found in freshwater and estuarine wetlands. Oligochaete density is positively correlated with soil organic matter and can be an indicator of soil development and wetland function in constructed and restored wetlands (Craft and Sacco 2003). In floodplain wetlands, burrowing by earthworms increases soil aeration and porosity, which facilitates plant establishment (Plum 2005).

Arthropoda

Arthropods include insects and crustaceans and are diverse and abundant in wetland soils.

Subphylum Insecta

Chironomid midges (fly larvae) are probably the most abundant insects inhabiting wetland soils; ants and termites are also relatively common (Bruskewitz 1981; Batzer and Wissinger 1996). Chironomids are sediment detritivores (Voshell 2002) that inhabit soil burrows and build protective tubes out of soil, detritus, or algae. Midge larvae can account for a large proportion of macroinvertebrate diversity in wetlands and provide an important food source for other invertebrates, birds, fish, and amphibians (Wissinger et al. 1999; Rosemond et al. 2001; Panatta et al. 2007). In mangrove mesocosms, burrows of chironomid larvae facilitated sediment oxidation to 12 cm below the surface (Kristensen and Alongi 2006). Some studies have shown that nitrification can be stimulated by chironomid larvae because of increased O_2 penetration into sediments, while others have shown suppression of nitrification by chironomids, possibly as a result of changes in nitrifier populations from ingestion of particle-associated bacteria by midges (Altmann et al. 2004). Termites act as ecosystem engineers in some wetlands. In tropical regions (e.g., the Pantanál do Mato Grosso, Brazil), termites build soil mounds above the saturated zone. These mounds enable tropical savanna (Cerrado) vegetation to establish, and the termite/vegetation interaction generates earth mounds that are analogous to marsh hummocks found in the Everglades and elsewhere (Ponce and Dacunha 1993).

Subphylum Crustacea

Crustaceans influence the chemistry and physical structure of wetland soils (Angeler et al. 2001; Batzer et al. 2006; Kristensen 2008). Burrowing by ocypodid (fiddler) crabs in tidal wetlands affects redox potential, mixes surface and subsurface soils, and moves labile and refractory compounds between upper and lower soil layers; in addition, crab feces can enrich sediments in ammonium (Montague 1982; Gribsholt et al. 2003; Cannicci et al. 2008). Burrowing promotes soil oxygenation and enables iron (rather than sulfate) reduction to be the dominant pathway of carbon mineralization in some salt marsh soils (Kristensen and Alongi 2006). In mangrove ecosystems, grazing by fiddler crabs (*Uca vocans*) and rhizosphere oxygenation by gray mangroves (*Avicennia marina*) facilitates iron oxidation and inhibits iron reduction in the upper sediment layers (Kristensen and Alongi 2006).

Crabs can improve the quality of detritus in mangrove ecosystems. Nitrogen enrichment of detritus by sesarmid (leaf-eating) crab feces leads to greater colonization by bacteria compared with unprocessed leaf litter (Cannicci et al. 2008). The crab *Heloecius cordiformis* (Milne Edwards) creates "hills" and redistributes soil particles between surface and belowground soil layers in Australian mangrove ecosystems (Warren and Underwood 1986). Wolfrath (1992) estimated that European fiddler crabs (*Uca tangeri*) rearranged 3000–6000 cm^3 sediments/($m^2 \cdot$ month) through burrowing activity in mudflats of Rio Formosa, Portugal (1992).

Parastacus, *Procambarus*, and *Orconectes* crayfish are important ecosystem engineers in many freshwater wetland habitats. Similar to crabs, burrowing by crayfish resuspends sediments and alters soil physiochemical properties, and these animals provide an important food source for birds and other vertebrates (Dorn and Volin 2009). Crayfish chambers are generally at the level of the water table; excavated sediments form "chimneys" at burrow entrances (Noro and Buckup 2010). *Parastacus defossus* burrows extended 1.15 m below the soil surface during the dry season in a Brazilian swamp and occupied approximately 1.4 m^2 of horizontal area, with multiple chimneys and tunnels per living chamber (Noro and Buckup 2010). Crayfish burrows can provide refugia for other invertebrates during drought or seasonal drawdown of the water table (Dietz-Brantley et al. 2002). Crayfish are omnivorous predators that can significantly alter food webs; their diet includes benthic, floating, and vascular vegetation, fish and amphibian eggs, gastropods and other smaller invertebrates (Lodge et al. 1994; Dorn and Wojdak 2004). The introduction of crayfish into areas where they are not native has led to disruptions in wetland hydrology and food webs (Cano and Ocete 1997; Hobbs et al. 1989).

Smaller crustaceans that inhabit wetland sediments include detritivorous ostracods (seed shrimp) (Taylor et al. 1999), amphipods (Anteau and Afton 2008), isopods (Smock and Harlowe 1983), and microcrustaceans. Microcrustaceans can comprise a substantial proportion of secondary production and overall invertebrate biomass in some wetlands and are important in transferring primary production to secondary consumers (Jenkins and Boulton 2003). In a perennial beaver-impounded wetland in Alabama, USA, benthic microcrustacean production, dominated by cladocerans and copepods, was approximately 13.5 g/m^2 year, compared with 2.4 g/m^2 year (dry-weight basis) for emerging insects (Lemke and Benke 2009).

Adaptations of Invertebrates to Wetland Soils

Wetland soil invertebrates are exposed to the challenges of oxygen deficiency, ionic stress, and toxic products of anaerobic metabolism including NH_3 and H_2S, acetic and other acids (Plum 2005). H_2S and NH_3 can diffuse across cell membranes and accumulate to lethal concentrations (Wang and Chapman 1999; Camargo and Alonso 2006). Various behavioral, anatomical, and metabolic adaptations to wetland soils occur among invertebrates (Table 5.3).

Mobile animals can escape anaerobic conditions by emigrating. Ant colonies self-assemble into living rafts, linking their bodies to form a floating "lifeboat" that they maintain

TABLE 5.3

Major Invertebrate Adaptations to Anaerobic Soils and Salinity

Adaptation to Flooding and Anaerobic Soils	Function	Suggested References
Burrow irrigation	Draw oxygenated water into habitat space	Gallon et al. (2008)
Vertical migration (e.g., climbing up vegetation)	Escape anaerobic conditions	Green et al. (2009)
Desiccation resistance	Eggs or spores remain dormant during dry conditions	Brock et al. (2003)
Emergence from burrows	Escape H_2S toxicity	Oseid and Smith (1974)
Rafting	Remain above water until soil is exposed; transport to new habitat	Mlot et al. (2011)
Adaptation to salinity		
Burrow into saturated zone	Avoid desiccation and maintain osmotic balance	Willmer (2006)
Regulate concentration of free amino acids in tissues	Maintain osmotic balance	Ferraris et al. (1994), Pequeux (1995)
Active transport of inorganic ions across cell walls	Maintain osmotic balance	Pequeux (1995)

until they reach dry ground, sometimes for months (Mlot et al. 2011). Gastropods, spiders, and other invertebrates climb vegetation to escape floodwaters (Green et al. 2009; Petillon et al. 2010). Some burrowing invertebrates (e.g., mayfly larvae) emerge from sediments to escape from toxic H_2S (Oseid and Smith 1974). Less-mobile benthic species have evolved physical and morphological adaptations, including parapodia or gills, enhanced respiratory pigments and plasticity in hemoglobin systems, regulation of oxygen uptake, and behaviors such as burrow irrigation (Table 5.3). Many polychaetes and gastropods have parapodia—specialized, paired appendages with movement—and respiration-related functions. Some parapodia contain gills or lamellae that increase the surface area for oxygen uptake (Kristensen 1983). The mantle epithelium of benthic and aquatic mollusks is highly vascularized to maximize oxygen uptake. In addition, some snails use their nuchal lobe to create a breathing siphon that they raise to the water surface (Santos et al. 1987). Some invertebrates synthesize hemoglobin in response to anoxia; diversity among hemoglobin molecules enables specialization for oxygen storage or transport (Weber 1980). The body fluid of chironomid larvae contains specialized hemoglobins and other pigments that have high affinity for oxygen (Schowalter 2011). Burrowing invertebrates use body movements to maintain circulation of water and favorable oxygen concentrations in their surroundings. Observed behaviors include crawling and turning, abdominal undulations, gill beating and rubbing, and sediment pushing (Gallon et al. 2008). In estuarine wetlands, *Uca* spp. seal themselves inside their air-filled burrows at rising tide (Havens 1990).

Invertebrate species in temporary wetlands such as vernal pools, which are typically dry in the summer, must be able to survive or resist desiccation. Mechanisms of desiccation resistance include aestivation (survival in an egg or spore stage in dry substrate) (Dietz-Brantley et al. 2002; Brock et al. 2003); retreat into terrestrial areas, below vegetation, or into soils (Green et al. 2009); and migration between ephemeral wetlands and permanent waterbodies (Hillman and Quinn 2002).

Over evolutionary time, invertebrates dispersed from marine habitats to terrestrial and freshwater environments and have evolved mechanisms for withstanding lower salt concentrations than are found in their environments of origin (Miller and Labandeira 2002). Some intertidal invertebrates burrow into sediments to reach the saturated zone, where they remain until tidal flooding resumes (Willmer 2006). Various estuarine invertebrates (e.g., some crustaceans and polychaetes) osmoregulate with inorganic ions and amino acids in response to changes in salinity (Ferraris et al. 1994; Pequeux 1995). Ion pumps transport waste products such as dissolved NH_4^+ and HCO_3^- across cell walls and exchange them for ions that are needed in higher concentrations, including Na^+ and Cl^- (Willmer 2006).

Vertebrates

Vertebrates affect wetland soils through their use of the habitat and their feeding activities. Animal-induced hydrological changes influence physical and biogeochemical processes in wetland soils. Some vertebrates (e.g., beavers and hippopotamuses) are considered wetland ecosystem engineers because of the magnitude of their influence. Herbivory by mammals including beavers, muskrats, capybara, and nutria affects the type and quantity of organic matter deposited in soils, as well as root growth and nutrient uptake (Ford and Grace 1998).

By building dams across low-order streams, beavers (*Castor canadensis* and *Castor fiber*) flood riparian areas and create lentic or standing-water habitats that develop into wetlands (Collen and Gibson 2001). Dam construction increases retention of sediments and organic matter, and over time the invertebrate fauna shifts from lotic (flowing water) taxa (e.g., Ephemeroptera, Trichoptera) to lentic, benthic species (e.g., many chironomids and oligochaetes) (Mcdowell and Naiman 1986). Decomposition rates decrease as the redox potential declines in beaver wetlands, which leads to accumulation of benthic organic matter and stimulation of anaerobic processes including methanogenesis (Naiman et al. 1986).

Hippos (*Hippopotamus amphibius*) control landscape morphology in African marshes and floodplains including the Okavango Delta (Botswana) (Eltringham 2001) and the Ngorongoro Crater wetlands (Tanzania). Hippos create distinctive topography in freshwater wetlands with soft sediments by wallowing and carving paths through vegetation. With annual changes in flooding and rainfall, wallows become mudflats or lagoons and dry hippo trails act as levees that alter sediment and water flows (Deocampo 2002). The large quantities of dung deposited by hippos stimulate primary productivity, and their bioturbation affects the oxygen status of sediments and the water column (Mosepele et al. 2009).

Crocodilians and large mammals influence wetland topography and vegetation in tropical and subtropical regions. Alligators displace vegetation as they move through it, and they maintain inundated wallows by excavating sediments and plants (Kushlan 1974; Campbell and Mazzotti 2004). Alligators also uproot significant amounts of vegetation to construct their nests (Joanen 1969). The world's largest rodent, the capybara (*Hydrochoerus hydrochaeris*) wallows and grazes in wetlands in tropical South America (e.g., the Pantanál) (Quintana and Rabinovich 1993). Herbivore grazing patterns may lead to increased plant productivity or enhanced nutrient cycling (the "grazing optimization hypothesis"), especially where coevolution between plants and herbivores has occurred (de Mazancourt and Loreau 2000). For example, in a flooded Venezuelan savanna, plants grazed by capybaras and other herbivores became enriched in nitrogen during the dry season to a significantly greater extent than the same species growing in plots that excluded grazers (Barreto and Herrera 1998).

Muskrats (*Ondatra zibethicus*) are important herbivores in freshwater and coastal wetlands (Allen and Hoffman 1984). Muskrats alter hydrology and soils by removing large quantities of vegetation and by constructing lodges and tunnels (De Szalay and Cassidy 2001). Their tracking through marshes is thought to redistribute and aerate surface soils (Connors et al. 2000). Another rodent, the coypu or nutria (*Myocastor coypus* Molina), causes extensive damage to brackish and freshwater wetlands outside its native range of South America. Nutria forage on roots and rhizomes of aquatic vegetation, which leaves the bare substrate vulnerable to erosion and subsidence (Carter et al. 1999). Although threatened by overhunting and habitat destruction in their home range, nutria populations in North America, Europe, and elsewhere have proven difficult to control (Carter and Leonard 2002; Guichon and Cassini 2005). Tens of thousands of acres of coastal wetlands in Louisiana have been lost to subsidence as a result of nutria herbivory (Marx 2004).

Livestock grazing influences physical and chemical characteristics of wetland soils. Large grazers compact soils, alter the plant community, and redistribute nutrients and seeds (Clement and Proctor 2009). The effects of grazing differ among wetland ecosystems and vary with the intensity of livestock use. Some wetlands, especially those found in grassland communities, evolved with large herbivores and may benefit from some level of grazing. In California, vernal pools grazed by cattle had a greater diversity of native plants and longer hydroperiod compared with pools from which cattle were excluded, which had greater cover of nonnative grasses and higher rates of primary productivity and transpiration (Marty 2005). Livestock also affect greenhouse gas fluxes. Methane emissions measured in spring-fed California wetlands and in a Tibetan alpine wetland were higher in grazed areas than in ungrazed plots, which might be explained by the release of CH_4 through cut plant stems, suppression of methane oxidation by soil compaction, or through effects of plant removal on hydrology (Allen-Diaz et al. 2004; Hirota et al. 2005). On the other hand, exclusion of grazers can lead to lower plant productivity and nitrogen uptake and higher rates of nitrous oxide production from increased availability of soil nitrogen (Jackson et al. 2006).

Conclusions

The distinguishing characteristics of wetland soils develop through biological and physical interactions. Microbial metabolism, plant growth and decomposition, herbivory, and bioturbation affect the soil structure and biogeochemical dynamics. The organisms that inhabit wetland soils have evolved physical and metabolic adaptations to periodic or continuous anaerobic conditions, and to salinity in estuarine ecosystems. The interplay between physical and biological elements in wetland soils generates the defining characteristics of these ecosystems.

Acknowledgments

I appreciate the time and input provided by Chris Craft during the preparation of this chapter. Thanks are also due to Maria Hernandez for reviewing a draft of the manuscript and to Jason Keller for helpful comments.

References

Allen, A. W. and Hoffman R. D. 1984. *Habitat Suitability Index Models: Muskrat*. U.S. Fish and Wildlife Service, Washington, DC. FWS/OBS-82/10.46.

Allen-Diaz, B., R. D. Jackson, J. W. Bartolome, K. W. Tate, and L. G. Oates. 2004. Long-term grazing study in spring-fed wetlands reveals management tradeoffs. *Calif. Agr.* 58: 144–148.

Altmann, D., P. Stief, R. Amann, and D. De Beer. 2004. Nitrification in freshwater sediments as influenced by insect larvae: Quantification by microsensors and fluorescence *in situ* hybridization. *Microb. Ecol.* 48: 145–153.

Amaranthus, M. P. 1998. The importance and conservation of ectomycorrhizal fungal diversity in forest ecosystems: Lessons from Europe and the Pacific Northwest. Gen. Tech. Rep. PNW-GTR-431. U.S. Department of Agriculture, Forest Service, Pacific Northwest Research Station, Portland, OR. 15 p.

Angeler, D. G., S. Sanchez-Carrillo, G. Garcia, and M. Alvarez-Cobelas. 2001. The influence of *Procambarus clarkii* (Cambaridae, Decapoda) on water quality and sediment characteristics in a Spanish floodplain wetland. *Hydrobiologia* 464: 89–98.

Anteau, M. J. and A. D. Afton. 2008. Amphipod densities and indices of wetland quality across the upper-Midwest, USA. *Wetlands* 28: 184–196.

Armstrong, W., S. H. F. W. Justin, P. M. Beckett, and S. Lythe. 1991. Root adaptation to soil waterlogging. *Aquat. Bot.* 39: 57–73.

Atlas, R. M. and Bartha, R. 1998. *Microbial Ecology: Fundamentals and Applications*. Addison Wesley Longman, Menlo Park, CA.

Balasooriya, W. K., K. Denef, J. Peters, N. E. C. Verhoest, and P. Boeckx. 2008. Vegetation composition and soil microbial community structural changes along a wetland hydrological gradient. *Hydrol. Earth Syst. Sci.* 12: 277–291.

Barreto, G. R. and E. A. Herrera. 1998. Foraging patterns of capybaras in a seasonally flooded savanna of Venezuela. *J. Trop. Ecol.* 14: 87–98.

Batzer, D. P., R. Cooper, and S. A. Wissinger. 2006. Wetland animal ecology. In D. P. Batzer, and R. R. Sharitz (Eds.) *Ecology of Freshwater and Estuarine Wetlands*. University of California Press, Berkeley, pp. 242–284.

Batzer, D. P. and R. R. Sharitz (Eds.) 2006. *Ecology of Freshwater and Estuarine Wetlands*. University of California Press, Berkeley.

Batzer, D. P. and S. A. Wissinger. 1996. Ecology of insect communities in nontidal wetlands. *Ann. Rev. Entomol.* 41: 75–100.

Bauer, C. R., C. H. Kellogg, S. D. Bridgham, and G. A. Lamberti. 2003. Mycorrhizal colonization across hydrologic gradients in restored and reference freshwater wetlands. *Wetlands* 23: 961–968.

Bever, J. D., P. A. Schultz, A. Pringle, and J. B. Morton. 2001. Arbuscular mycorrhizal fungi: More diverse than meets the eye, and the ecological tale of why. *Bioscience* 51: 923–931.

Beylich, A. and U. Graefe. 2002. Annelid coenoses of wetlands representing different decomposer communities. In G. Broll, W. Merbach, and E. M. Pfeiffer (Eds.) *Wetlands in Central Europe: Soil Organisms, Soil Ecological Processes and Trace Gas Emissions.* Springer, Berlin, pp. 1–10.

Blokhina, O., E. Virolainen, and K. V. Fagerstedt. 2003. Antioxidants, oxidative damage and oxygen deprivation stress: A review. *Ann. Bot.-London* 91: 179–194.

Blom, C. W. P. M. 1999. Adaptations to flooding stress: From plant community to molecule. *Plant Biol.* 1: 261–273.

Bonkowski, M., C. Villenave, and B. Griffiths. 2009. Rhizosphere fauna: The functional and structural diversity of intimate interactions of soil fauna with plant roots. *Plant Soil* 321: 213–233.

Boon, P. I. 2006. Biogeochemistry and bacterial ecology of hydrologically dynamic wetlands. In D. P. Batzer, and R. R. Sharitz (Eds.) *Ecology of Freshwater and Estuarine Wetlands*. University of California Press, Berkeley, pp. 115–176.

Botella, M. A., A. Rosado, R. A. Bressan, and P. M. Hasegawa. 2005. Plant adaptive responses to salinity stress. In M. A. Jenks and P. M. Hasegawa (Eds.) *Plant Abiotic Stress*. Blackwell Publishing Ltd, Oxford, UK, pp. 37–70.

Bowden, W. B. 1987. The Biogeochemistry of nitrogen in freshwater wetlands. *Biogeochemistry* 4: 313–348.

Bradley, P. M. and J. T. Morris. 1990. Influence of oxygen and sulfide concentration on nitrogen uptake kinetics in *Spartina alterniflora*. *Ecology* 71: 282–287.

Bridgham, S. D., H. Cadillo-Quiroz, J. K. Keller, and Q. Zhuang. 2013. Methane emissions from wetlands: Biogeochemical, microbial, and modeling perspectives from local to global scales. *Glob. Change Biol.* 19: 1325–1346.

Brinson, M. M., A. E. Lugo, and S. Brown. 1981. Primary productivity, decomposition and consumer activity in freshwater wetlands. *Ann. Rev. Ecol. Syst.* 12: 123–162.

Brix, H. 1997. Do macrophytes play a role in constructed treatment wetlands? *Water Sci. Technol.* 35: 11–17.

Brock, M. A., D. L. Nielsen, R. J. Shiel, J. D. Green, and J. D. Langley. 2003. Drought and aquatic community resilience: The role of eggs and seeds in sediments of temporary wetlands. *Freshwater Biol.* 48: 1207–1218.

Bruskewitz, J. W. 1981. Wetland ants: Internal mound temperature and humidity preferences; location and shape of mounds as adaptations to a wetland environment. *Trans. Wis. Acad. Sci.* 69: 21–25.

Buesing, N. and M. O. Gessner. 2006. Benthic bacterial and fungal productivity and carbon turnover in a freshwater marsh. *Appl. Environ. Microbiol.* 72: 596–605.

Burgin, A. J. and S. K. Hamilton. 2007. Have we overemphasized the role of denitrification in aquatic ecosystems? A review of nitrate removal pathways. *Front. Ecol. Environ.* 5: 89–96.

Camargo, J. A. and A. Alonso. 2006. Ecological and toxicological effects of inorganic nitrogen pollution in aquatic ecosystems: A global assessment. *Environ. Int.* 32: 831–849.

Campbell, M. R. and F. J. Mazzotti. 2004. Characterization of natural and artificial alligator holes. *Southeast. Nat.* 3: 583–594.

Cannicci, S., D. Burrows, S. Fratini, T. J. Smith, J. Offenberg, and F. Dahdouh-Guebas. 2008. Faunal impact on vegetation structure and ecosystem function in mangrove forests: A review. *Aquat. Bot.* 89: 186–200.

Cano, E. and M. E. Ocete. 1997. Population biology of red swamp crayfish, *Procambarus clarkii* (Girard, 1852) in the Guadalquivar River marshes, Spain. *Crustaceana* 70: 553–561.

Caraco, N., J. Cole, and G. E. Likens. 1990. A comparison of phosphorus immobilization in sediments of freshwater and coastal marine systems. *Biogeochemistry* 9: 277–290.

Carter, J., A. L. Foote, and L. A. Johnson-Randall. 1999. Modeling the effects of nutria (*Myocastor coypus*) on wetland loss. *Wetlands* 19: 209–219.

Carter, J. and B. P. Leonard. 2002. A review of the literature on the worldwide distribution, spread of, and efforts to eradicate the coypu (*Myocastor coypus*). *Wildlife Soc. Bull.* 30: 162–175.

Carvalho, L. M., I. Cacador, and M. A. Martins-Loucao. 2001. Temporal and spatial variation of arbuscular mycorrhizas in salt marsh plants of the Tagus estuary (Portugal). *Mycorrhiza* 11: 303–309.

Castignetti, D. and T. C. Hollocher. 1982. Nitrogen redox metabolism of a heterotrophic, nitrifying-denitrifying *Alcaligenes* sp. from soil. *Appl. Environ. Microbiol.* 44: 923–928.

Chaban, B., S. Y. M. Ng, and K. F. Jarrell. 2006. Archaeal habitats—From the extreme to the ordinary. *Can. J. Microbiol.* 52: 73–116.

Chen, R. and R. R. Twilley. 1999. A simulation model of organic matter and nutrient accumulation in mangrove wetland soils. *Biogeochemistry* 44: 93–118.

Chmura, G. L., S. C. Anisfeld, D. R. Cahoon, and J. C. Lynch. 2003. Global carbon sequestration in tidal, saline wetland soils. *Global Biogeochem. Cy.* 17(4): 1111, doi: 10.1029/2002GB001917.

Clement, B. and M. C. F. Proctor. 2009. Ecological dynamics I: Vegetation as bioindicator and dynamic community. In E. Maltby, and T. Barker (Eds.) *The Wetlands Handbook*. Wiley-Blackwell Publishing, Oxford, UK, pp. 282–303.

Clément, J. C., G. Pinay, and P. Marmonier. 2002. Seasonal dynamics of denitrification along topohydrosequences in three different riparian wetlands. *J. Environ. Qual.* 31: 1025–1037.

Clément, J. C., S. Shrestha, J. G. Ehrenfeld, and P. R. Jaffé. 2005. Ammonium oxidation coupled to dissimilatory reduction of iron under anaerobic conditions in wetland soils. *Soil Biol. Biochem.* 37: 2323–2328.

Collen, P. and R. J. Gibson. 2001. The general ecology of beavers (*Castor* spp.), as related to their influence on stream ecosystems and riparian habitats, and the subsequent effects on fish—A review. *Rev. Fish Biol. Fisher.* 10: 439–461.

Colmer, T. D. 2003. Long-distance transport of gases in plants: A perspective on internal aeration and radial oxygen loss from roots. *Plant Cell Environ.* 26: 17–36.

Connors, L. M., E. Kiviat, P. M. Groffman P.M, and R. S. Ostfeld. 2000. Muskrat (*Ondatra zibethicus*) disturbance to vegetation and potential net nitrogen mineralization and nitrification rates in a freshwater tidal marsh. *Am. Midl. Nat.* 143: 53–63.

Craft, C. B., S. W. Broome, E. D. Seneca, and W. J. Showers. 1988. Estimating sources of soil organic matter in natural and transplanted estuarine marshes using stable isotopes of carbon and nitrogen. *Estuar. Coast. Shelf Sci.* 26: 633–642.

Craft, C. B. and W. P. Casey. 2000. Sediment and nutrient accumulation in floodplain and depressional freshwater wetlands of Georgia, USA. *Wetlands* 20: 323–332.

Craft, C. B. and C. J. Richardson. 1993. Peat accretion and nitrogen, phosphorus, and organic carbon accumulation in nutrient-enriched and unenriched Everglades peatlands. *Ecol. Appl.* 3: 446–458.

Craft, C. B. and J. Sacco. 2003. Long-term succession of benthic infauna communities on constructed *Spartina alterniflora* marshes. *Mar. Ecol. Prog. Ser.* 257: 45–58.

Crawford, R. M. M. and R. Braendle. 1996. Oxygen deprivation stress in a changing environment. *J. Exp. Bot.* 47: 145–159.

Cronk, J. K. and M. S. Fennessy. 2001. *Wetland Plants: Biology and Ecology*. Lewis Publishers, Boca Raton, FL.

Crow, S. E. and R. K. Wieder. 2005. Sources of CO_2 emission from a northern peatland: Root respiration, exudation, and decomposition. *Ecology* 86: 1825–1834.

Daleo, P., J. Alberti, A. Canepuccia, M. Escapa, E. Fanjul, B. R. Silliman, M. D. Bertness, and O. Iribarne. 2008. Mycorrhizal fungi determine salt-marsh plant zonation depending on nutrient supply. *J. Ecol.* 96: 431–437.

Davelaar, D. 1993. Ecological significance of bacterial polyphosphate metabolism in sediments. *Hydrobiologia* 253: 179–192.

de Mazancourt, C. and M. Loreau. 2000. Grazing optimization, nutrient cycling, and spatial heterogeneity of plant-herbivore interactions: Should a palatable plant evolve? *Evolution* 54: 81–92.

Deocampo, D. M. 2002. Sedimentary structures generated by *Hippopotamus amphibius* in a lake-margin wetland, Ngorongoro Crater, Tanzania. *Palaios* 17: 212–217.

De Szalay, F. A. and W. Cassidy. 2001. Effects of muskrat (*Ondatra zibethicus*) lodge construction on invertebrate communities in a great lakes coastal wetland. *Am. Midl. Nat.* 146: 300–310.

Dietz-Brantley, S. E., B. E. Taylor, D. P. Batzer, and A. E. DeBiase. 2002. Invertebrates that aestivate in dry basins of Carolina bay wetlands. *Wetlands* 22: 767–775.

Dise, N. B. 2009. Biogeochemical dynamics III: The critical role of carbon in wetlands. In E. Maltby, and T. Barker (Eds.) *The Wetlands Handbook*. Wiley-Blackwell Publishing, Oxford, UK, pp. 249–265.

Dorn, N. J. and J. C. Volin. 2009. Resistance of crayfish (Procambarus spp.) populations to wetland drying depends on species and substrate. *J. N. Am. Benthol. Soc.* 28: 766–777.

Dorn, N. J. and J. M. Wojdak. 2004. The role of omnivorous crayfish in littoral communities. *Oecologia* 140: 150–159.

Ehrenfeld, J. G., B. Ravit, and K. Elgersma. 2005. Feedback in the plant-soil system. *Annu. Rev. Env. Resour.* 30: 75–115.

Eitam, A., C. Norena, and L. Blaustein. 2004. Microturbellarian species richness and community similarity among temporary pools: Relationships with habitat properties. *Biodivers. Conserv.* 13: 2107–2117.

Ellery, W. N., T. S. McCarthy, and N. D. Smith. 2003. Vegetation, hydrology, and sedimentation patterns on the major distributary system of the Okavango Fan, Botswana. *Wetlands* 23: 357–375.

Eltringham, S. K. 2001. Hippopotamuses. In D. Macdonald (Ed.) *The Encyclopedia of Mammals*. Andromeda Oxford Limited, UK.

Fageria, N. K., V. C. Baligar, and R. B. Clark. 2002. Micronutrients in crop production. *Adv. Agron.* 77: 185–268.

Fahn, A. 1988. Secretory tissues in vascular plants. *New Phytol.* 108: 229–258.

Ferraris, J. D., K. Fauchald, and B. Kensley. 1994. Physiological responses to fluctuation in temperature or salinity in invertebrates: Adaptations of *Alpheus viridari* (Decapoda, Crustacea), *Terebellids parva* (Polychaeta) and *Golfinigia cylindrata* (Sipunculida) to the mangrove habitat. *Mar. Biol.* 120: 397–406.

Ford, M. A. and J. B. Grace. 1998. Effects of vertebrate herbivores on soil processes, plant biomass, litter accumulation and soil elevation changes in a coastal marsh. *J. Ecol.* 86: 974–982.

Gallon, C., L. Hare, and A. Tessier. 2008. Surviving in anoxic surroundings: How burrowing aquatic insects create an oxic microhabitat. *J. N. Am. Benthol. Soc.* 27: 570–580.

Gessner, M. O., V. Gulis, K. A. Kuehn, E. Chauvet, and K. Suberkropp. 2007. Fungal decomposers of plant litter in aquatic ecosystems. In C. P. Kubicek, and I. S. Druzhinina (Eds.) *The Mycota*, Vol. IV, *Environmental and Microbial Relationships*. Springer-Verlag, Berlin, pp. 301–324.

Gill, A. M. and P. B. Tomlinson. 1977. Studies on the growth of red mangrove (*Rhizophora mangle* L.). Part 4: The adult root system. *Biotropica* 9: 145–155.

Gillett, D. J., A. F. Holland, and D. A. Sanger. 2007. On the ecology of oligochaetes: Monthly variation of community composition and environmental characteristics in two South Carolina tidal creeks. *Estuar. Coast.* 30: 238–252.

Gorham, E. 1991. Northern peatlands: Role in the carbon cycle and probable responses to climatic warming. *Ecol. Appl.* 1: 182–195.

Govedich, F. R., B. A. Bain, W. E. Moser, S. R. Gelder, R. W. Davies, and R. O. Brinkhurst. 2010. Annelida (Clitellata): Oligochaeta, Branchiobdellida, Hirudinida, and Acanthobdellida. In J. H. Thorp and A. P. Covich (Eds.) *Ecology and Classification of North American Freshwater Invertebrates*. Academic Press (Elsevier), Amsterdam, pp. 385–436.

Green, J., A. Reichelt-Brushett, and S. W. L. Jacobs. 2009. Investigating gastropod habitat associations in saltmarsh. *Wetlands* 25: 25–37.

Gribsholt, B., J. E. Kostka, and E. Kristensen. 2003. Impact of fiddler crabs and plant roots on sediment biogeochemistry in a Georgia saltmarsh. *Mar. Ecol. Prog. Ser.* 259: 237–251.

Guichon, M. L. and M. H. Cassini. 2005. Population parameters of indigenous populations of *Myocastor coypus*: The effect of hunting pressure. *Acta Theriol.* 50: 125–132.

Gulis, V., K. Kuehn, and K. Suberkropp. 2006. The role of fungi in carbon and nitrogen cycles in freshwater ecosystems. In G. M. Gadd (Ed.) *Fungi in Biogeochemical Cycles*. Cambridge University Press, Cambridge, UK, pp. 404–435.

Gutknecht, J. L. M., R. M. Goodman, and T. C. Balser. 2006. Linking soil process and microbial ecology in freshwater wetland ecosystems. *Plant Soil* 289: 17–34.

Hajibagheri, M. A., A. R. Yeo, and T. J. Flowers. 1985. Salt tolerance in *Suaeda maritima* (L.) Dum. Fine structure and ion concentration in the apical region of roots. *New Phytol.* 99: 331–344.

Harley, J. L. 1989. The significance of mycorrhiza. *Mycol. Res.* 92: 129–139.

Hasegawa, P. M., R. A. Bressan, J. K. Zhu, and H. J. Bohnert. 2000. Plant cellular and molecular responses to high salinity. *Annu. Rev. Plant Physiol. Plant Mol. Biol.* 51: 463–499.

Havens, K. J. 1990. Animals of the intertidal mud flats (a glance). Virginia Institute of Marine Science Technical Report No. 90–1. Gloucester Point, VA.

Herbert, R. A. and D. B. Nedwell. 1990. Role of environmental factors in regulating nitrate respiration in intertidal sediments. In N. P. Revsbech, and E. Sorensen (Eds.) *Denitrification in Soil and Sediment*. Plenum Press, New York, pp. 77–90.

Hernandez, M. E. and W. J. Mitsch. 2007. Denitrification in created riverine wetlands: Influence of hydrology and season. *Ecol. Eng.* 30: 78–88.

Hillman, T. J. and G. P. Quinn. 2002. Temporal changes in macroinvertebrate assemblages following experimental flooding in permanent and temporary wetlands in an Australian floodplain forest. *River Res. Appl.* 18: 137–154.

Hirota, M., Y. Tang, Q. Hu, T. Kato, S. Hirata, W. Mo, G. Cao, and S. Mariko. 2005. The potential importance of grazing to the fluxes of carbon dioxide and methane in an alpine wetland on the Qinghai-Tibetan Plateau. *Atmos. Environ.* 39: 5255–5259.

Hobbie, J. E., E. A. Hobbie, H. Drossman, M. Conte, J. C. Weber, J. Shamhart, and M. Weinrobe. 2009. Mycorrhizal fungi supply nitrogen to host plants in Arctic tundra and boreal forests: N-15 is the key signal. *Can. J. Microbiol.* 55: 84–94.

Hobbs, H. H. III, J. P. Jass, and J. V. Huner. 1989. A review of global grayfish introductions with particular emphasis on two North American species (Decapoda, Cambaridae). *Crustaceana* 56: 299–316.

Hofferle, S., G. W. Nicol, L. Pal, J. Hacin, J. I. Prosser, and I. Mandic-Mulec. 2010. Ammonium supply rate influences archaeal and bacterial ammonia oxidizers in a wetland soil vertical profile. *FEMS Microbiol. Ecol.* 74: 302–315.

Holguin, G., P. Vazquez, and Y. Bashan. 2001. The role of sediment microorganisms in the productivity, conservation, and rehabilitation of mangrove ecosystems: An overview. *Biol. Fert. Soils* 33: 265–278.

Howarth, R. W., R. Marino, J. Lane, and J. J. Cole. 1988. Nitrogen fixation in freshwater, estuarine and marine ecosystems 1. Rates and importance. *Limnol. Oceanogr.* 33: 669–687.

Howes, B. L., J. W. H. Dacey, and G. M. King. 1984. Carbon flow through oxygen and sulfate reduction pathways in salt marsh sediments. *Limnol. Oceanogr.* 29: 1037–1051.

Hyun, J. H., A. C. Smith, and J. E. Kostka. 2007. Relative contributions of sulfate- and iron (III) reduction to organic matter mineralization and process controls in contrasting habitats of the Georgia saltmarsh. *Appl. Geochem.* 22: 2637–2651.

Ipsilantis, I. and D. M. Sylvia. 2007. Abundance of fungi and bacteria in a nutrient-impacted Florida wetland. *Appl. Soil Ecol.* 35: 272–280.

Jackson, M. B. and W. Armstrong. 1999. Formation of aerenchyma and the processes of plant ventilation in relation to soil flooding and submergence. *Plant Biol.* 1: 274–287.

Jackson, R. D., B. Allen-Diaz, L. G. Oates, and K. W. Tate. 2006. Spring-water nitrate increased with removal of livestock grazing in a California oak savanna. *Ecosystems* 9: 254–267.

Jenkins, K. M. and A. J. Boulton. 2003. Connectivity in a dryland river: Short-term aquatic microinvertebrate recruitment following floodplain inundation. *Ecology* 84: 2708–2723.

Jenks, M. A. and P. M. Hasegawa (Eds.) 2005. *Plant Abiotic Stress*. Blackwell Publishing Ltd, Oxford, UK.

Jensen, K. T. and L. D. Kristensen. 1990. A field experiment on competition between *Corophium volutator* (Pallas) and *Corophium arenarium* Crawford (Crustacea: Amphipoda): Effects on survival reproduction and recruitment. *J. Exp. Mar. Biol. Ecol.* 137: 1–24.

Joanen, T. 1969. Nesting ecology of alligators in Louisiana. *Proc. Annu. Conf. Southeast. Assoc. Game and Fish Comm.* 24: 141–151.

Jones, D. L., C. Nguyen, and R. D. Finlay. 2009. Carbon flow in the rhizosphere: Carbon trading at the soil–root interface. *Plant Soil* 321: 5–33.

Jordan, T. E., J. C. Cornwell, W. R. Boynton, and J. T. Anderson. 2008. Changes in phosphorus biogeochemistry along an estuarine salinity gradient: The iron conveyer belt. *Limnol. Oceanogr.* 53: 172–184.

Keeley, J. E. 1980. Endomycorrhizae influence growth of black-gum seedlings in flooded soils. *Am. J. Bot.* 67: 6–9.

Kelaher, B. P., J. S. Levinton, and J. M. Hoch. 2003. Foraging by the mud snail, *Ilyanassa obsoleta* (Say), modulates spatial variation in benthic community structure. *J. Exp. Mar. Biol. Ecol.* 292: 139–157.

Kelso, B. H. L., R. V. Smith, and R. J. Laughlin. 1999. Effects of carbon substrates on nitrite accumulation in freshwater sediments. *Appl. Environ. Microbiol.* 65: 61–66.

Kirk, G. J. D. and H. J. Kronzucker. 2005. The potential for nitrification and nitrate uptake in the rhizosphere of wetland plants: A modelling study. *Ann. Bot.* 96: 639–646.

Kletzin, A. 2007. General characteristics and important model organisms. In R. Cavicchioli (Ed.) *Archaea: Molecular and Cellular Biology*. ASM Press, Washington, DC, pp. 14–92.

Kolasa, J. and S. Tyler. 2010. Turbellarians and nemertea. In J. H. Thorp, and A. P. Covich (Eds.) *Ecology and Classification of North American Freshwater Invertebrates*. Academic Press (Elsevier), Amsterdam, pp. 143–161.

Koncalova, H. 1990. Anatomical adaptations to waterlogging in roots of wetland graminoids—Limitations and drawbacks. *Aquat. Bot.* 38: 127–134.

Konhauser, K. O. 1998. Diversity of bacterial iron mineralization. *Earth Sci. Rev.* 43: 91–121.

Koop-Jakobsen, K. and A. E. Giblin. 2010. The effect of increased nitrate loading on nitrate reduction via denitrification and DNRA in salt marsh sediments. *Limnol. Oceanogr.* 55: 789–802.

Koretsky, C. M., C. M. Moore, K. L. Lowe, C. Meile, T. J. Dichristina, and P. van Cappellen. 2003. Seasonal oscillation of microbial iron and sulfate reduction in saltmarsh sediments (Sapelo Island, GA, USA). *Biogeochemistry* 64: 179–203.

Kostka, J. E., A. Roychoudhury, and P. van Cappellen. 2002. Rates and controls of anaerobic microbial respiration across spatial and temporal gradients in saltmarsh sediments. *Biogeochemistry* 60: 49–76.

Kristensen, E. 1983. Ventilation and oxygen uptake by 3 species of *Nereis* (Annelida: Polychaeta). 1. Effects of hypoxia. *Mar. Ecol. Prog. Ser.* 12: 289–298.

Kristensen, E. 2008. Mangrove crabs as ecosystem engineers; with emphasis on sediment processes. *J. Sea Res.* 59: 30–43.

Kristensen, E. and D. M. Alongi. 2006. Control by fiddler crabs (*Uca vocans*) and plant roots (*Avicennia marina*) on carbon, iron, and sulfur biogeochemistry in mangrove sediment. *Limnol. Oceanogr.* 51: 1557–1571.

Kristensen, E., F. O., Andersen, N. Holmboe, M. Holmer, and N. Thongtham. 2000. Carbon and nitrogen mineralization in sediments of the Bangrong mangrove area, Phuket, Thailand. *Aquat. Microb. Ecol.* 22: 199–213.

Kurz, H. and D. Demaree. 1934. Cypress buttresses and knees in relation to water and air. *Ecology* 15: 36–41.

Kushlan, J. A. 1974. Observations on role of American alligator (*Alligator Mississippiensis*) in Southern Florida Wetlands. *Copeia* 1974: 993–996.

Laanbroek, H. J. 2010. Methane emission from natural wetlands: Interplay between emergent macrophytes and soil microbial processes. A mini-review. *Ann. Bot.* 105: 141–153.

Langley, J. A. and B. A. Hungate. 2003. Mycorrhizal controls on belowground litter quality. *Ecology* 84: 2302–2312.

Lawrence, B. A. and J. B. Zedler. 2011. Formation of tussocks by sedges: Effects of hydroperiod and nutrients. *Ecol. Appl.* 21: 1745–1759.

Lemke, A. M. and A. C. Benke. 2009. Spatial and temporal patterns of microcrustacean assemblage structure and secondary production in a wetland ecosystem. *Freshwater Biol.* 54: 1406–1426.

Leonard, L. A., P. A. Wren, and R. L. Beavers. 2002. Flow dynamics and sedimentation in *Spartina alterniflora* and *Phragmites australis* marshes of the Chesapeake Bay. *Wetlands* 22: 415–424.

Levin, L. A. 1984. Life history and dispersal patterns in a dense infaunal polychaete assemblage: Community structure and response to disturbance. *Ecology* 65: 1185–1200.

Levinsh, G. 2006. Biological basis of biological diversity: Physiological adaptations of plants to heterogeneous habitats along a sea coast. *Acta U. Latviensis* 710: 53–79.

Liu, D. Y., W. X. Ding, Z. J. Jia, and Z. C. Cai. 2011. Relation between methanogenic archaea and methane production potential in selected natural wetland ecosystems across China. *Biogeosciences* 8: 329–338.

Lodge, D. M., M. W. Kershner, J. E. Aloi, and A. P. Covich. 1994. Effects of an omnivorous crayfish (*Orconectes rusticus*) on a freshwater littoral food web. *Ecology* 75: 1265–1281.

Lovley, D. R. and E. J. P. Phillips. 1986. Organic matter mineralization with reduction of ferric iron in anaerobic sediments. *Appl. Environ. Microb.* 51: 683–689.

Mackenzi, F. T. and R. M. Garrels. 1966. Chemical mass balance between rivers and oceans. *Am. J. Sci.* 264: 507–525.

Mansfield, S. D. and F. Barlocher. 1993. Seasonal variation of fungal biomass in the sediment of a salt marsh in New Brunswick. *Microb. Ecol.* 26: 37–45.

Marty, J. T. 2005. Effects of cattle grazing on diversity in ephemeral wetlands. *Conserv. Biol.* 19: 1626–1632.

Marx, J. M., E. Mouton, and G. Linscombe. 2004. *Nutria Harvest Distribution 2003–2004 and a Survey of Nutria Herbivory Damage in Coastal Louisiana in 2004*. Fur and Refuge Division Louisiana Department of Wildlife and Fisheries, 1–45.

Mcdowell, D. M. and R. J. Naiman. 1986. Structure and function of a benthic invertebrate stream community as influenced by beaver (*Castor Canadensis*). *Oecologia* 68: 481–489.

Megonigal, J. P., M. E. Hines, and P. T. Visscher. 2004. Anaerobic metabolism: Linkages to trace gases and aerobic processes. In W. H. Schlesinger (Ed.) *Biogeochemistry*. Elsevier-Pergamon, Oxford, UK, pp. 317–424.

Mendelssohn, I. A. and D. P. Batzer. 2006. Abiotic constraints for wetland plants and animals. In D. P. Batzer and R. R. Sharitz (Eds.) *Ecology of Freshwater and Estuarine Wetlands*. University of California Press, Berkeley, pp. 82–114.

Mendelssohn, I. A., K. L. McKee, and W. H. J. Patrick. 1981. Oxygen deficiency in *Spartina alterniflora* roots: Metabolic adaptation to anoxia. *Science* 214: 439–441.

Merritt, R. W., J. R. Wallace, M. J. Higgins, M. K. Alexander, M. B. Berg, W. T. Morgan, K. W. Cummins, and B. Vandeneeden. 1996. Procedures for the functional analysis of invertebrate communities of the Kissimmee River-floodplain ecosystem. *Fla. Scientist* 59: 216–274.

Middleton, B. A. and K. L. McKee. 2001. Degradation of mangrove tissues and implications for peat formation in Belizean island forests. *J. Ecol.* 89: 818–828.

Miller, M. F. and C. Labandeira. 2002. Slow crawl across the salinity divide: Delayed colonization of freshwater ecosystems by invertebrates. *Geol. Soc. Am. Today* 12: 4–10.

Miller, S. P. and J. D. Bever. 1999. Distribution of arbuscular mycorrhizal fungi in stands of the wetland grass *Panicum hemitomon* along a wide hydrologic gradient. *Oecologia* 119: 586–592.

Mitsch, W. J. and J. G. Gosselink. 2000. *Wetlands*. John Wiley, New York.

Mlot, N. J., C. A. Tovey, and D. L. Hu. 2011. Fire ants self-assemble into waterproof rafts to survive floods. *Proc. Nat. Acad. Sci. USA* 108: 7669–7673.

Mohamed, D. J. and J. B. H. Martiny. 2011. Patterns of fungal diversity and composition along a salinity gradient. *ISME J.* 5: 379–388.

Montague, C. L. 1982. The influence of fiddler crab burrows and burrowing on metabolic processes in salt marsh sediments. In V. S. Kennedy (Ed.) *Estuarine Comparisons*. Academic Press, New York, pp. 283–301.

Moran, M. A. and R. E. Hodson. 1992. Contributions of three subsystems of a freshwater marsh to total bacterial secondary productivity. *Microb. Ecol.* 24: 161–170.

Moran, M. A., A. E. Maccubbin, R. Benner, and R. E. Hodson. 1987. Dynamics of microbial biomass and activity in five habitats of the Okefenokee Swamp ecosystem, Georgia USA. *Microb. Ecol.* 14: 203–218.

Mosepele, K., P. B. Moyle, G. S. Merron, D. R. Purkey, and B. Mosepele. 2009. Fish, floods, and ecosystem engineers: Aquatic conservation in the Okavango Delta, Botswana. *Bioscience* 59: 53–64.

Munns, R. and M. Tester. 2008. Mechanisms of salinity tolerance. *Ann. Rev. Plant Biol.* 59: 651–681.

Naidoo G., K. L. Mckee, and I. A. Mendelssohn. 1992. Anatomical and metabolic responses to waterlogging and salinity in *Spartina alterniflora* and *S. patens* (Poaceae). *Am. J. Bot.* 79: 765–770.

Naidoo, G. and R. Rughunanan. 1990. Salt tolerance in the succulent coastal halophyte. *Journal of Experimental Botany* 41: 497–502.

Naiman, R. J., J. M. Melillo, and J. E. Hobbie. 1986. Ecosystem alteration of boreal forest streams by beaver (*Castor canadensis*). *Ecology* 67: 1254–1269.

Newell, S. Y., M. A. Moran, R. Wicks, and R. E. Hodson. 1995. Productivities of microbial decomposers during early stages of decomposition of leaves of a freshwater sedge. *Freshwater Biol.* 34: 135–148.

Nguyen, C. 2003. Rhizodeposition of organic C by plants: Mechanisms and controls. *Agronomie* 23: 375–396.

Noro, C. K. and L. Buckup. 2010. The burrows of *Parastacus defossus* (Decapoda: Parastacidae), a fossorial freshwater crayfish from southern Brazil. *Zoologia* 27: 341–346.
Nyman, J. A., R. D. DeLaune, H. H. Roberts, and W. H. Patrick Jr. 1993. Relationship between vegetation and soil formation in a rapidly submerging coastal marsh. *Mar. Ecol.-Prog. Ser.* 96: 269–279.
Ogburn, R. M. and E. J. Edwards. 2010. The ecological water-use strategies of succulent plants. *Adv. Bot. Res.* 55: 179–225.
Oseid, D. M. and L. L. J. Smith. 1974. Factors influencing acute toxicity estimates of hydrogen sulfide to fresh water invertebrates. *Water Res.* 8: 739–746.
Otte, M. L. 2001. What is stress to a wetland plant? *Environ. Exp. Bot.* 46: 195–202.
Padgett, D. E. and D. A. Celio. 1990. A newly discovered role for aerobic fungi in anaerobic salt marsh soils. *Mycologia* 82: 791–794.
Paerl, H. W. and J. L. Pinckney. 1996. A mini-review of microbial consortia: Their roles in aquatic production and biogeochemical cycling. *Microb. Ecol.* 31: 225–247.
Panatta, A., C. Stenert, E. M. dos Santos, and L. Maltchik. 2007. Diversity and distribution of chironomid larvae in wetlands in southern Brazil. *J. Kansas Entomol. Soc.* 80: 229–242.
Parniske, M. 2008. Arbuscular mycorrhiza: The mother of plant root endosymbioses. *Nat. Rev. Microbiol.* 6: 763–775.
Pazinato, J. M., E. N. Paulo, L. W. Mendes, R. F. Vazoller, and S. M. Tsai. 2010. Molecular characterization of the archaeal community in an Amazonian wetland soil and culture-dependent isolation of methanogenic Archaea. *Diversity* 2: 1026–1047.
Pequeux, A. 1995. Osmotic regulation in crustaceans. *J. Crustacean Biol.* 15: 1–60.
Petillon, J., K. Lambeets, W. Montaigne, J. P. Maelfait, and D. Bonte. 2010. Habitat structure modified by an invasive grass enhances inundation withstanding in a salt-marsh wolf spider. *Biol. Invasions* 12: 3219–3226.
Pezeshki, S. R., J. H. Pardue, and R. D. Delaune. 1993. The influence of soil oxygen deficiency on alcohol dehydrogenase activity, root porosity, ethylene production and photosynthesis in *Spartina patens*. *Environ. Exp. Bot.* 33: 565–573.
Plum, N. 2005. Terrestrial invertebrates in flooded grassland: A literature review. *Wetlands* 25: 721–737.
Poinar, G. O. J. 2010. Nematoda and nematomorpha. In J. H. Thorp and A. P. Covich (Eds.) *Ecology and Classification of North American Freshwater Invertebrates*. Academic Press (Elsevier), Amsterdam, pp. 237–276.
Ponce, V. M. and C. N. Dacunha. 1993. Vegetated earthmounds in tropical savannas of Central Brazil: A synthesis with special feference to the Pantanal do Mato Grosso. *J. Biogeogr.* 20: 219–225.
Ponnamperuma, F. N. 1972. The chemistry of submerged soils. *Adv. Agron.* 24: 29–96.
Prakasam, T. B. S. 1984. Denitrification. *J. Water Pollut. Con. F.* 56: 582–586.
Quintana, R. D. and J. E. Rabinovich. 1993. Assessment of capybara (*Hydrochoerus hydrochaeris*) populations in the wetlands of Corrientes, Argentina. *Wetl. Ecol. Manag.* 2: 223–230.
Raghoebarsing, A. A., A. Pol, K.T. van de Pas-Schoonen, et al. 2006. A microbial consortium couples anaerobic methane oxidation to denitrification. *Nature* 440: 918–921.
Rains, D. W. 1972. Salt transport by plants in relation to salinity. *Ann. Rev. Plant Physiol.* 23: 367–388.
Reddy, K. R. and E. M. D'Angelo. 1994. Soil processes regulating water quality in wetlands. In W. J. Mitsch (Ed.) *Global Wetlands: Old World and New*. Elsevier Science, Amsterdam, pp. 309–324.
Reddy, K. R., E. M. D'Angelo, and W. G. Harris. 2000. Biogeochemistry of wetlands. In M. Sumner (Ed.) *Handbook of Soil Science*. CRC Press, Boca Raton, FL, pp. G89–G119.
Reddy, K. R. and R. D. DeLaune. 2008. *Biogeochemistry of Wetlands: Science and Applications*. CRC Press, Boca Raton, FL.
Reddy, K. R., R. H. Kadlec, E. Flaig, and P. M. Gale. 1999. Phosphorus retention in streams and wetlands: A review. *Crit. Rev. Environ. Sci. Technol.* 29: 83–146.
Rich, S. M., M. Ludwig M, and T. D. Colmer. 2012. Aquatic adventitious root development in partially and completely submerged wetland plants *Cotula coronopifolia* and *Meionectes brownii*. *Ann. Bot.* 110: 405–414.

Richardson, C. J. and P. Vaithiyanathan. 2009. Biogeochemical dynamics II: Cycling and storage of phosphorus in wetlands. In E. Maltby and T. Barker (Eds.) *The Wetlands Handbook*. Wiley-Blackwell Publishing, Oxford, UK, pp. 228–248.

Ridge, I. 1987. Ethylene and growth control in amphibious plants. In R. M. M. Crawford (Ed.) *Plant Life in Aquatic and Amphibious Habitats*. Blackwell Scientific, Oxford, UK, pp. 53–76.

Roden, E. E. and Q. Jin. 2011. Thermodynamics of microbial growth coupled to metabolism of glucose, ethanol, short-chain organic acids, and hydrogen. *Appl. Environ. Microb.* 77: 1907–1909.

Roden, E. E. and R. G. Wetzel. 1996. Organic carbon oxidation and suppression of methane production by microbial Fe(III) oxide reduction in vegetated and unvegetated freshwater wetland sediments. *Limnol. Oceanogr.* 41: 1733–1748.

Rosemond, A. D., C. M. Pringle, A. Ramirez, and M. J. Paul. 2001. A test of top-down and bottom-up control in a detritus-based food web. *Ecology* 82: 2279–2293.

Rouse, G. W. 2001. *Annelida (Segmented Worms).* In eLS. John Wiley & Sons Ltd, Chichester. http://www.els.net [doi: 10.1038/npg.els.0001599].

Rublee, P. A. 1982. Bacteria and microbial distribution in estuarine sediments. In V. S. Kennedy (Ed.) *Estuarine Comparisons*, Academic Press, New York, pp. 159–182.

Rutting, T., P. Boeckx, C. Mueller, and L. Klemedtsson. 2011. Assessment of the importance of dissimilatory nitrate reduction to ammonium for the terrestrial nitrogen cycle. *Biogeosciences* 8: 1779–1791.

Santos, C. A., C. H. S. Penteado, and E. G. Mendes. 1987. The respiratory responses of an amphibious snail *Pomacea lineata* (Spix, 1827) to temperature and oxygen tension variations. *Comp. Biochem. Physiol.* 86A: 409–416.

Schimel, J. P. and J. Bennett. 2004. Nitrogen mineralization: Challenges of a changing paradigm. *Ecology* 85: 591–602.

Schlesinger, W. H. 1997. *Biogeochemistry: An Analysis of Global Change*, 2nd ed. Academic Press, San Diego, CA.

Schowalter, T. D. 2011. *Insect Ecology: An Ecosystem Approach*, 3rd ed. Academic Press (Elsevier), San Diego, CA.

Schussler, E. E. and D. J. Longstreth. 1996. Aerenchyma develops by cell lysis in roots and cell separation in leaf petioles in *Sagittaria lancifolia* (Alismataceae). *Am. J. Bot.* 83: 1266–1273.

Seitzinger, S., J. A. Harrison, J. K. Bohlke, A. F. Bouwman, R. Lowrance R, B. Peterson, C. Tobias, and G. Van Drecht. 2006. Denitrification across landscapes and waterscapes: A synthesis. *Ecol. Appl.* 16: 2064–2090.

Sengupta, A. and S. Chaudhuri. 2002. Arbuscular mycorrhizal relations of mangrove plant community at the Ganges river estuary in India. *Mycorrhiza* 12: 169–174.

Shapiro, J. A. 1988. Bacteria as multicellular organisms. *Sci. Am.* 258: 82–89.

Shoun, H. and T. Tanimoto. 1991. Denitrification by the fungus *Fusarium oxysporum* and involvement of cytochrome P-450 in the respiratory nitrite reduction. *J. Biol. Chem.* 266: 11078–11082.

Siddiqui, Z. A. and J. Pichtel. 2008. Mycorrhizae: An overview. In Z. A. Siddiqui, M. S. Akhtar, and K. Futai (Eds.) *Mycorrhizae: Sustainable Agriculture and Forestry*. Springer, Dordrecht, The Netherlands, pp. 1–35.

Sigg, L. 2000. Redox potential measurements in natural waters: Significant concepts and problems. In H. D. Schulz, W. R. Fischer, J. Bottcher, and W. H. M. Duijnisveld (Eds.) *Redox: Fundamentals, Processes and Applications*. Springer, Berlin, Germany.

Silliman, B. R. and M. D. Bertness. 2002. A trophic cascade regulates salt marsh primary production. *Proc. Nat. Acad. Sci. USA* 99: 10500–10505.

Smock, L. A. and K. L. Harlowe. 1983. Utilization and processing of freshwater wetland macrophytes by the detritivore *Asellus forbesi*. *Ecology* 64: 1556–1565.

Snowden, R. E. D. and B. D. Wheeler. 1993. Iron toxicity to fen plant-species. *J. Ecol.* 81: 35–46.

Su, R., R. N. Lohner, K. A. Kuehn, R. Sinsabaugh, and R. K. Neely. 2007. Microbial dynamics associated with decomposing *Typha angustifolia* litter in two contrasting Lake Erie coastal wetlands. *Aquat. Microb. Ecol.* 46: 295–307.

Sundareshwar, P. V., J. T. Morris, P. J. Pellechia, H. J. Cohen, D. E. Porter, and B. C. Jones. 2001. Occurrence and ecological implications of pyrophosphate in estuaries. *Limnol. Oceanogr.* 46: 1570–1577.

Tabak, H. H. and W. B. Cooke. 1968. Effects of gaseous environments on growth and metabolism of fungi. *Bot. Rev.* 34: 126–252.

Taylor, B. E., D. A. Leeper, M. A. McClure, and A. E. DeBiase. 1999. Carolina bays: Ecology of aquatic invertebrates and perspectives on conservation. In D. P. Batzer, R. B. Rader, and S. A. Wissinger (Eds.) *Invertebrates in Freshwater Wetlands of North America: Ecology and Management*. John Wiley & Sons, Inc., New York, NY, pp. 167–196.

Teal, J. M. 1962. Energy flow in the salt marsh ecosystem of Georgia. *Ecology* 43: 614–624.

Thormann, M. N., R. S. Currah, and S. E. Bayley. 1999. The mycorrhizal status of the dominant vegetation along a peatland gradient in southern boreal Alberta, Canada. *Wetlands* 19: 438–450.

Thorp, J. H., A. P. Covich, and W. W. Dimmick. 2010. Introduction to invertebrates of inland waters. In J. H. Thorp and A. P. Covich (Eds.) *Ecology and Classification of North American Freshwater Invertebrates*. Academic Press (Elsevier), San Diego, CA, pp. 1–23.

Tiedje, J. M. 1988. Ecology of denitrification and dissimilatory nitrate reduction to ammonium. In A. J. B. Zehnder (Ed.) *Biology of Anaerobic Microorganisms*. John Wiley & Sons, New York, NY, pp. 179–244.

Tiner, R. W. 1999. Wetland indicators: A guide to wetland identification, delineation, classification, and mapping. Lewis Publishers, Boca Raton, FL.

Tonouchi, A. 2009. Isolation and characterization of a novel facultative anaerobic filamentous fungus from Japanese rice field soil. *Int. J. Microbiol.* 2009: 571383.

Turner, S. D., J. P. Amon, R. M. Schneble, and C. F. Friese. 2000. Mycorrhizal fungi associated with plants in groundwater fed wetlands. *Wetlands* 20: 200–204.

Van der Valk, A. G. 2006. *Biology of Freshwater Wetlands*. Oxford University Press, Oxford, UK.

Varnell, L. M. 1998. The relationship between inundation history and baldcypress stem form in a Virginia floodplain swamp. *Wetlands* 18: 176–183.

Vepraskas, M. J. 1992. *Redoximorphic Features for Identifying Aquic Conditions*. Tech. Bull. 301. NC Agric. Exp. Sta., Raleigh, NC.

Visser, E. J. W., T. D. Colmer, C. W. P. M. Blom, and L. A. C. J. Voesenek. 2000. Changes in growth, porosity, and radial oxygen loss from adventitious roots of selected mono- and dicotyledonous wetland species with contrasting types of aerenchyma. *Plant Cell Environ.* 23: 1237–1245.

Volkmar, K. M., Y. Hu, and H. Steppuhn. 1998. Physiological responses of plants to salinity: A review. *Can. J. Plant Sci.* 78: 19–27.

Voshell, J. R. 2002. *A Guide to Common Freshwater Invertebrates of North America*. McDonald & Woodward Publishers, Blacksburg, VA.

Vymazal, J. 2007. Removal of nutrients in various types of constructed wetlands. *Sci. Total Environ.* 380: 48–65.

Wainwright, M. 1988. Metabolic diversity of fungi in relation to growth and mineral cycling in soil: A review. *Trans. Brit. Mycol. Soc.* 90: 159–170.

Wainwright, M. 1992. The impact of fungi on environmental biogeochemistry. In G. C. Carroll and M. V. Wicklow (Eds.) *The Fungal Community: Its Organization and Role in the Ecosystem*, 2nd ed. Marcel Dekker, Inc., New York, NY, pp. 601–618.

Waisel, Y., A. Eshel, and M. Agami. 1986. Salt balance of leaves of the mangrove *Avicennia marina*. *Physiol. Plantarum* 67: 67–72.

Wang, F. and P. M. Chapman. 1999. Biological implications of sulfide in sediment: A review focusing on sediment toxicity. *Environ. Toxicol. Chem.* 18: 2526–2532.

Warren, J. H. and A. J. Underwood. 1986. Effects of burrowing crabs on the topography of mangrove swamps in New South Wales. *J. Exp. Mar. Biol. Ecol.* 102: 223–235.

Warwick, R. M. and R. Price. 1979. Ecological and metabolic studies on free-living nematodes from an estuarine mud-flat. *Estuar. Coast. Mar. Sci.* 9: 257–271.

Waughman, G. J. and D. J. Bellamy. 1980. Nitrogen fixation and the nitrogen balance in peatland ecosystems. *Ecology* 61: 1185–1198.

Weber, K. A., L. A. Achenbach, and J. D. Coates. 2006a. Microorganisms pumping iron: Anaerobic microbial iron oxidation and reduction. *Nat. Rev. Microbiol.* 4: 752–764.
Weber, K. A., M. M. Urrutia, P. F. Churchill, R. K. Kukkadapu, and E. E. Roden. 2006b. Anaerobic redox cycling of iron by freshwater sediment microorganisms. *Environ. Microbiol.* 8: 100–113.
Weber, R. E. 1980. Functions of invertebrate hemoglobins with special reference to adaptations to environmental hypoxia. *Am. Zool.* 20: 79–102.
Weishampel, P. A. and B. L. Bedford. 2006. Wetland dicots and monocots differ in colonization by arbuscular mycorrhizal fungi and dark septate endophytes. *Mycorrhiza* 16: 495–502.
Weston, N. B. and S. B. Joye. 2005. Temperature-driven decoupling of key phases of organic matter degradation in marine sediments. *Proc. Nat. Acad. Sci. USA* 102: 17036–17040.
Weston, N. B., W. P. Porubsky, V. A. Samarkin, M. Erickson, S. E. Macavoy, and S. B. Joye. 2006. Porewater stoichiometry of terminal metabolic products, sulfate, and dissolved organic carbon and nitrogen in estuarine intertidal creek-bank sediments. *Biogeochemistry* 77: 375–408.
Whigham, D. F. 2009. Global distribution, diversity and human alterations of wetland resources. In E. Maltby and T. Barker (Eds.) *The Wetlands Handbook*. Wiley-Blackwell Publishing, Oxford, UK, pp. 43–64.
White, J. R. and K. R. Reddy. 2009. Biogeochemical dynamics I: Nitrogen cycling in wetlands. In E. Maltby and T. Barker (Eds.) *The Wetlands Handbook*. Wiley-Blackwell Publishing, Oxford, UK, pp. 213–227.
Wieser, W. and J. Kanwisher. 1961. Ecological and physiological studies on marine nematodes from a small salt marsh near Woods Hole, Massachusetts. *Limnol. Oceanogr.* 6: 262–270.
Willmer, P. 2006. *Osmoregulation in Invertebrates*. In eLS. John Wiley & Sons Ltd, Chichester. http://www.els.net [doi: 10.1038/npg.els.0003646].
Wissinger, S. A. 1999. Ecology of wetland invertebrates: Synthesis and applications for conservation and management. In D. P. Batzer, R. B. Rader, and S. A. Wissinger (Eds.) *Invertebrates in Freshwater Wetlands of North America: Ecology and Management*. John Wiley & Sons, Inc., New York, NY, pp. 1043–1086.
Wissinger, S. A., A. J. Bohonak, H. H. Whiteman, and W. S. Brown. 1999. Subalpine wetlands in Colorado: Habitat permanence, salamander predation, and invertebrate communities. In D. P. Batzer, R. B. Rader, and S. A. Wissinger (Eds.) *Invertebrates in Freshwater Wetlands of North America: Ecology and Management*. John Wiley & Sons, Inc., New York, NY, pp. 757–790.
Wolfe, B. E., P. A. Weishampel, and J. N. Klironomos. 2006. Arbuscular mycorrhizal fungi and water table affect wetland plant community composition. *J. Ecol.* 94: 905–914.
Wolfrath, B. 1992. Burrowing of the fiddler crab *Uca tangeri* in the Ria Formosa in Portugal and its influence on sediment structure. *Mar. Ecol.-Prog. Ser.* 85: 237–243.
Wright, R. B., B. G. Lockaby, and M. R. Walbridge. 2001. Phosphorus availability in an artificially flooded southeastern floodplain forest soil. *Soil Sci. Soc. Am. J.* 65: 1293–1302.
Wu, H. Y., X. X. Li, L. B. Shi, Z. H. Wang, and F. Y. Ma. 2008. Distribution of nematodes in wetland soils with different distance from the Bohai sea. *Plant Soil Environ.* 54: 359–366.
Zhou, Z., N. Takaya, A. Nakamura, M. Yamaguchi, K. Takeo, and H. Shoun. 2002. Ammonia fermentation, a novel anoxic metabolism of nitrate by fungi. *J. Biol. Chem.* 277: 1892–1896.
Zhu, G., S. Wang, Y. Wang, C. Wang, N. Risgaard-Petersen, M. S. Jetten, and C. Yin. 2011. Anaerobic ammonia oxidation in a fertilized paddy soil. *ISME J.* 5: 1905–1912.

6

Soil Organic Matter

Jason K. Keller and Cassandra A. Medvedeff

CONTENTS

Introduction

The Soil Science Society of America defines soil organic matter (SOM) as "the organic fraction of the soil exclusive of undecayed plant and animal residues" (Soil Science Society of America 2008). SOM, here used interchangeably with "humus," is built upon a chemical backbone of carbon, hydrogen, and oxygen, but also contains other important elements including organic nitrogen, organic phosphorus, and organic forms of a number of trace elements. Ultimately, SOM is derived from senescent plant material, either produced autochthonously or produced allochthonously and captured by an ecosystem, which has been at least partially degraded by the decomposer community. As decomposers mineralize organic carbon substrates to fuel their metabolism, they simultaneously mineralize the organic nutrients necessary for subsequent plant and microbial growth. Thus, SOM represents an important link between plant and microbial activities within ecosystems.

On a global scale, ~1500 Pg (Pg = 10^{15} g) of carbon is stored in the SOM contained within the top 1 m of soil (Batjes 1996; and Amundson 2001), with an estimated 2300 Pg of carbon stored globally in plant detritus and SOM to a depth of 3 m (Jobbágy and Jackson 2000). The carbon stored in the upper 1 m of soil alone is two orders of magnitude larger than CO_2 released annually from fossil fuel burning (Schlesinger and Bernhardt 2013). Wetland soils store a disproportionately large fraction of this carbon—approximately 500 Pg (Bridgham et al. 2006, and references cited therein)—primarily as a result of flooded or saturated conditions, which limit oxygen availability and result in fundamentally different dynamics of decomposition than in generally aerobic terrestrial soils. The same anaerobic (i.e., low oxygen) conditions that lead to accumulation of SOM in wetland soils, however, result in the production of the important greenhouse gas methane (CH_4) by microbial decomposition, and wetlands play a critical role in the global cycling of this gas (Bartlett and Harriss 1993; Tian et al. 2010; Bridgham et al. 2013). Thus, understanding SOM cycling and processing within wetland soils has important implications in the context of the global carbon cycle and global climate change.

In this chapter, we begin by defining SOM with a focus on the various classes of chemical compounds included in the modern definition of SOM. Next, we define organic soils and explore their global distribution. Our goal is not to provide an extensive discussion of organic soil taxonomy and as such we limit our discussion to only the broadest categories of organic soils. We include a brief overview of some of the most important chemical and physical properties of organic soils and discuss the ecology of SOM with a particular emphasis on anaerobic processing of SOM that is unique to wetland environments. Finally, we provide examples of the connections between wetland SOM and human activities.

Soil Organic Matter

SOM is composed of products resulting from the partial decomposition of plant and animal residues (Figure 6.1; Soil Science Society of America 2008). While not explicit in the definition, for the purposes of this chapter, SOM includes both solid-phase organic matter as well

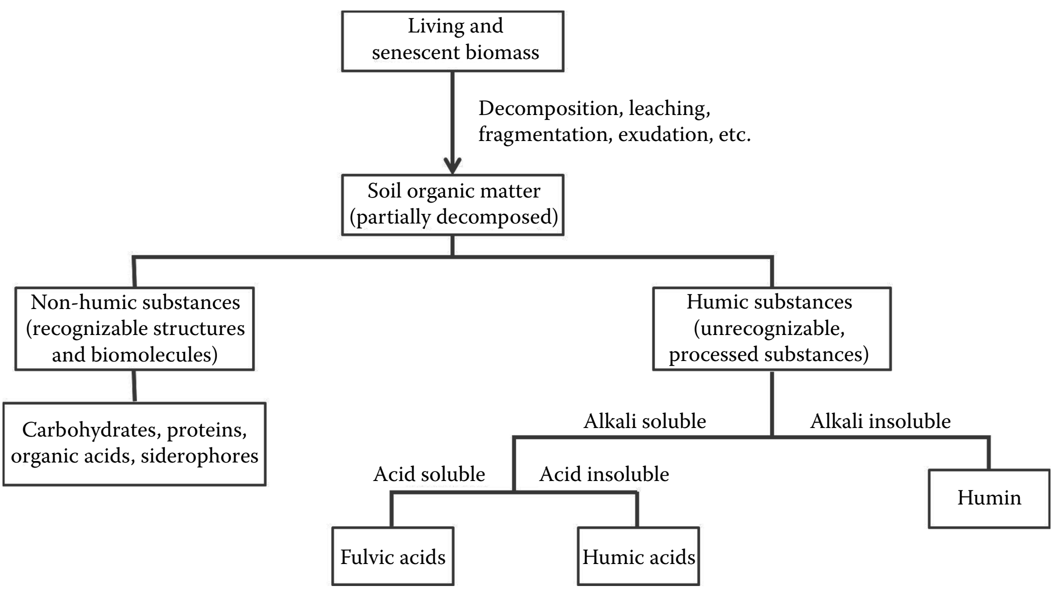


FIGURE 6.1
Soil organic matter is composed of partially decomposed plant and animal residues. Typically, soil organic matter is divided into both non-humic and humic substances. Humic substances are traditionally further characterized by alkali and acid solubility.

as dissolved organic matter contained in soil porewater of flooded or saturated wetland soils. Undecayed plant and animal materials are explicitly excluded from the definition of SOM, as are living roots and rhizomes and all other belowground macroorganisms, including plants and animals (Soil Science Society of America 2008). In contrast, organic materials associated with both living and dead microorganisms are often included as a fraction of SOM. In reality, however, the distinction between undecayed materials and SOM is hard to define as the transition from litter residue to SOM represents a continuum (e.g., Melillo et al. 1989). For example, leaching can very quickly transfer soluble compounds from leaf litter into SOM without significantly altering the physical structure of that litter. Even the distinction between living macroorganisms and SOM is likely to be a bit blurred. For example, living roots facilitate complex carbon dynamics in the rhizosphere (Lynch and Whipps 1990; Bodelier et al. 2006; Jones et al. 2009) with root exudates, secretions, and root turnover being derived from living roots but released into SOM.

At the coarsest level, SOM is typically divided into non-humic and humic substances (Figure 6.1; Stevenson 1994). Non-humic compounds are characterized by recognizable chemical structures and typically include the biomolecules necessary to sustain the living soil biomass (Sposito 2008), and comprise about 20%–30% of the total SOM (Brady and Weil 2008). Biomolecules are both added and removed from SOM during the initial stages of decomposition of plant or animal residues as a result of microbial metabolism and immobilization. Common biomolecules found in SOM include carbohydrates, organic acids, proteins, and lipids associated with plant and microbial biomass and decomposer processes (Reddy and DeLaune 2008, Sposito 2008).

A majority of carbohydrates found in wetland soils are likely to be polysaccharides of high molecular weight, including starch, cellulose, and hemicellulose that are common components of vegetation (Reddy and DeLaune 2008). Lower molecular weight carbohydrates, including some organic acids, are less important in terms of total SOM mass, but can be important metabolic intermediates in the dissolved organic matter pool (e.g., Duddleston et al. 2002; Drake et al. 2009; Ye et al. 2014). A number of complex aromatic compounds, including lignins and tannins, are also important components of vegetation that generally limit decomposition and can accumulate as SOM. Unlike polysaccharides, these compounds do not have regular components and linkages and are thus not considered polymers. Research suggests many microbial groups may be inhibited by phenolic compounds (Mellegård et al. 2009; Stalheim et al. 2009; Bragazza et al. 2012), thus altering the rate of organic matter degradation and subsequent C greenhouse gas (CO_2 and CH_4) emissions. Microbial processing of plant-derived phenolics may also be an important pathway of humus formation.

Proteins (polymers of amino acids linked by peptide bonds) are important molecules in living cells and contribute up to 50% of microbial biomass, up to 20% of living plant biomass, and can contribute between 2% and 20% of wetland and peat SOM (Reddy and DeLaune 2008). Even when they contribute a small percentage to total SOM, proteins are an important pool of organic nitrogen and are important in soil nitrogen cycling. There is an increasing body of evidence demonstrating that many plants are capable of directly taking up organic nitrogen in the form of amino acids monomers (Näsholm et al. 2009).

Sposito (2008) includes siderophores—low-molecular-mass compounds synthesized by bacteria, plants, and fungi for Fe(III) acquisition—as a class of biomolecules in SOM. Siderophores have been shown to increase Fe(III) solubility under flooded conditions in a variety of wetland soils (e.g., Kolditz et al. 2009; Lipson et al. 2012) and may be an important and dynamic component of wetland SOM.

In contrast to the recognizable biomolecules of non-humic SOM, humic substances are traditionally characterized as high-molecular-weight substances resulting from the processing of biomolecules in the soil (Figure 6.1; Table 6.1; Soil Science Society of America 2008). Central to this definition is the concept that these compounds are not biopolymers synthesized directly to sustain the life cycles of soil biomass, but rather are generated as by-products of microbial metabolism (Sposito 2008). These materials frequently contribute up to 80% of soil humus and up to 50% of dissolved organic matter (Sposito 2008) and are considered to be largely refractory in terms of microbial decomposition. Despite their importance, there is little consensus on the particular mechanisms of humic substance formation, but it is generally assumed to involve contributions from both plant and microbial residues (Stevenson 1994), with a growing appreciation for the role of microbial residues (Miltner et al. 2012).

Historically, humic substances are further divided into classes based on solubility in alkaline and acidic solutions (Figure 6.1; Stevenson 1994). Fulvic acids are defined as the organic materials that are extractable from soils using a dilute basic solution and that remain soluble in acidic solutions (i.e., fulvic acids, are soluble in both acids and bases). Humic acids are defined as the organic materials that are extractable from soils using a dilute basic solution and that are insoluble in acidic solutions with a pH of 1–2. The remaining non-alkali-extractable portion of SOM is defined as humin (Swift 1996).

While the classes of humic substances (i.e., fulvic acids, humic acids, and humin) are operationally defined based on differential solubility, it was often assumed that these classes represented distinct chemical materials in SOM. Recent advances have challenged this view. The refined view of humic substances no longer considers these materials as distinct classes of macromolecules. Instead, the new view shows soil humic substances to be better described as supramolecular associations of diverse low molecular mass biomolecules forming dynamic associations based on hydrogen bonds and hydrophobic interactions (Sutton and Sposito 2005; Kelleher and Simpson 2006; Lehmann et al. 2008; Sposito 2008). This changing paradigm does not rule out the existence of humic substances

TABLE 6.1

Key Definitions Related to Soil Organic Matter

Soil organic matter	The organic fraction of the soil exclusive of undecayed plant and animal residues.
Humic substances	A series of relatively high-molecular-weight, yellow to black colored, organic substances formed by secondary synthesis reactions in soils. The term is used in a generic sense to describe the colored material or its fractions obtained on the basis of solubility characteristics. These materials are distinctive to soil environments in that they are dissimilar to the biopolymers of microorganisms and higher plants (including lignin).
Organic soils	A soil that classifies as a Histosol (see below).
Organic soil materials	Soil materials that have 18% or more organic carbon if the mineral fraction has 60% or more clay, or 12% organic carbon if the mineral fraction has no clay or has proportional amounts of organic carbon for intermediate clay contents.
Histosols	Organic soils that have organic soil materials in more than half of the upper 80 cm, or that are of any thickness if overlying rock or fragmental materials that have interstices filled with organic soil materials.
Bog	An organic-accumulating wetland that has no significant inflows or outflows and supports acidophilic mosses, particularly *Sphagnum*.
Fen	A peat accumulating wetland that receives some drainage from surrounding mineral soils and usually supports marshlike vegetation. These areas are richer in nutrients and less acidic than bogs. The soils under fens are peats (Histosols) if the fen has been present for a while.

Source: Adapted from Soil Science Society of America. 2008. *Glossary of Soil Science Terms*. Soil Science Society of America, Inc., Madison, WI, and Soil Survey Staff. 2010. *Keys to Soil Taxonomy*, 11th ed. USDA-Natural Resources Conservation Service, Washington, DC.

Note: Definitions are listed in the order that they appear in this chapter.

as an important component of SOM resulting from biological processing and condensation of plant residues, but rather challenges their existence as distinct chemical compounds. It should also be noted that a great deal of this work has focused on terrestrial humic materials, and a synthetic review of wetland humics has yet to be completed.

The changing view of soil humics has been part of a larger reevaluation of the stability of the different components of SOM. It has traditionally been assumed that non-humic biomolecules represent labile organic matter and are susceptible to rapid microbial processing and thus have short residence times in larger SOM pools. In contrast, soil humics have been viewed as chemically recalcitrant and were thought to contribute the majority of long-lived SOM pools. New research has challenged these views and reinforces the importance of environmental and biological processes in soil as key controls over SOM stability (e.g., Kleber 2010; Schmidt et al. 2011; Dungait et al. 2012). Included in the list of controls on SOM stability are microbial community structure and composition, freeze–thaw dynamics, fire dynamics, and the mineral protection of SOM from microbial degradation (Schmidt et al. 2011). Given the large amount of SOM stored in wetland soils and its potential to influence the global carbon cycle and global change, an evaluation of these new perspectives is particularly important.

Organic Soils

Defining Organic Soils

Organic soils are soils in which organic materials are present in greater abundance than mineral materials (Table 6.1; Soil Science Society of America 2008). Organic soil materials in this definition are defined based on the content of organic carbon in a soil as well as the clay content of the soil (Figure 6.2; Chapter 1). The USDA further allows for the use of

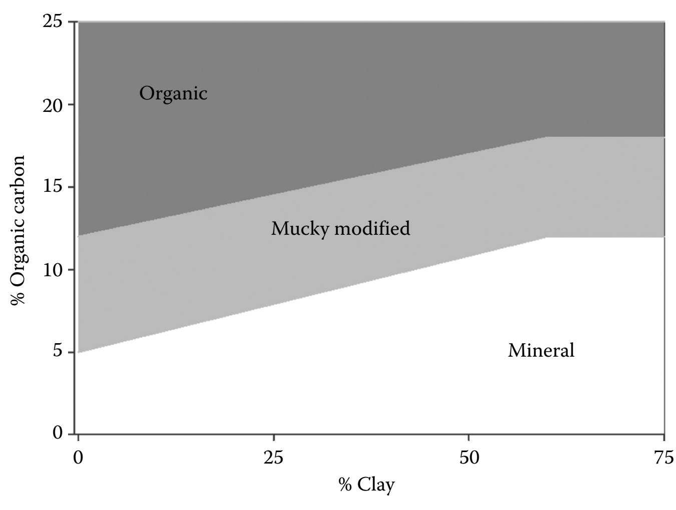


FIGURE 6.2
Definition of organic, mucky modified, and mineral soils based on percent organic carbon and clay content. (Modified from Natural Resources Conservation Service. 2010. *Field Indicators of Hydric Soils in the United States, Version 7.0.* Natural Resources Conservation Service, Washington, DC.)

the modifier "mucky" to describe the texture of mineral soils that have significant organic carbon content but less than that required for organic soils (Figure 6.2; Chapter 1).

Within the US Soil Classification System (Chapter 1, Soil Survey Staff 2010), organic soils are classified within the order Histosols. Histosols are organic soils with at least a 40-cm thick organic soil material layer at the surface (Soil Survey Staff 2010). For saturated wetland soils, the presence of >40 cm of surface organic material generally classifies a soil as a Histosol (Table 6.1; Soil Survey Staff 2010). Histosols (Chapter 10) are further divided into five suborders: Folists, Wassists, Fibrists, Saprists, and Hemists (Figure 6.3; Soil Survey Staff 2010). Folists are organic soils that are not permanently saturated but that are not artificially drained. Wassists are subaqueous Histosols that are submerged and thus exhibit a positive surface water potential for more than 21 h each day. The remaining suborders—Fibrists, Hemists, and Saprists—are all saturated under regular field conditions (unless artificially drained) but differ in terms of degree of decomposition (see further discussion below; Table 6.2). Fibrists contain the least decomposed soil organic materials

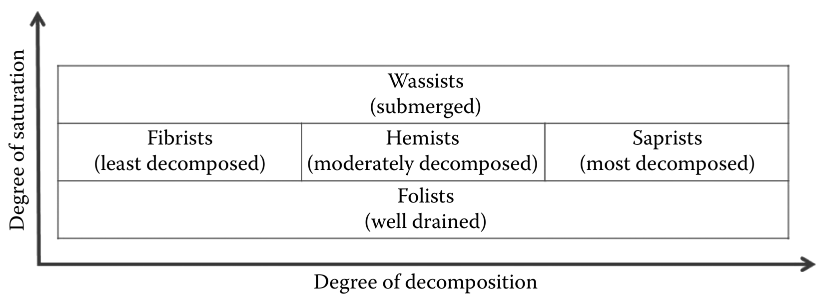


FIGURE 6.3
Five suborders of histosols are defined based on the degree of saturation and degree of decomposition.

TABLE 6.2

Degree of Decomposition of Organic Soils Based on Fiber Volume and Pyrophosphate Color Analysis Methods

	Fibric	Hemic	Sapric
Fiber volume	≥75% (by volume) fiber after rubbing, or ≥40% (by volume) fiber after rubbing and pyrophosphate color as described below	Between 17% and 40% (by volume) fiber after rubbing	<17% (by volume) fiber after rubbing and pyrophosphate color as described below
Pyrophosphate color analysis	Light Colors color value/chroma 8/1, 8/2, 8/3 7/1, 7/2		Dark Colors color value/chroma 7/4, 7/6, 7/8 6/3, 6/4, 6/6, 6/8 5/2, 5/3, 5/4, 5/6, 5/8 4/all chroma 3/all chroma 2/all chroma

Note: Color values and chroma for the pyrophosphate color analysis refer to the 7.5YR or 10YR Munsell color charts (see Figure 1.3). Color values represent lightness and chromas indicate the strength or departure from a neutral of the same lightness. All values are taken from Soil Survey Staff (2010).

(fibric material); Saprists have the most decomposed organic material (sapric material); and Hemists have intermediately decomposed organic materials (hemic material).

In addition to Histosols, many soils in the order Gelisols are also classified as organic soils. Gelisols are soils of very cold climates characterized by permafrost within the upper 2 m (Chapter 10). Histels within the Gelisols are generally defined as soils with organic materials that are >80% by volume of the upper 50 cm or that extend to a dense or bedrock layer in shallower soils (Soil Survey Staff 2010). The current US Soil Taxonomy System also allows for the identification of high organic content in otherwise mineral soils through the use of the histic epipedon designation. Epipedons are soil horizons that form at or near the surface, and histic epipedons are organic horizons that are generally 20–40 cm thick (Soil Survey Staff 2010). Thus, the fundamental difference between a mineral soil with a histic epipedon and an organic soil is the thickness of the surface organic layer.

Histosols are commonly referred to as peatlands, which share the characteristic of at least 40 cm of surface organic matter based on the definition provided above. Within this broad definition of peatlands an organic-rich wetlands, natural historians, and ecologists have long recognized many different types of peatlands in the landscape, including bogs and fens (Table 6.1). However, as pointed out by Bridgham et al. (1996), "Exactly how to define bog and fen can cause heated debate among otherwise mild-mannered wetland scientists."

Much of the challenge in peatland classification comes about because of co-occurring and often co-correlated gradients in alkalinity, pH, nutrient availability, and vegetation among different peatland types (Bridgham et al. 1996). Kolka et al. (Chapter 10) focus on the ombrogenous–minerogenous gradient in their treatment of Histosols. End members of this gradient range from ombrogenous bogs, which receive only atmospheric inputs, to minerogenous rich fens that have strong connections to the surface and/or groundwater inputs. Other peatland types, for example, poor fens, fall between these end members in their degree in hydrological connectivity. While this gradient is based on hydrology, the resulting classifications are related to vegetation and biogeochemistry, which differ among peatland types. The central role of hydrology in classifying peatlands is also reflected in the definitions used by the Soil Science Society of America (Table 6.1; 2008).

Global Distribution of Wetland Organic Matter

Given their high organic content, peatland ecosystems store the vast majority of wetland SOM on the global scale. A global accounting of organic matter stored in peatlands typically relies on estimates of peatland area and soil carbon density (mass of carbon per area to a given depth, typically expressed as Mg C ha^{-1}).

Similar to the overall global wetland distribution, peatlands follow a nonrandom distribution with high concentrations in both northern latitudes and the tropics (Chapter 10). Peatland distribution is particularly skewed towards northern latitudes (Figure 6.4; Yu et al. 2010). Recent estimates of peatland area suggest that northern peatlands may cover between 3.5×10^6 km^2 (Tarnocai et al. 2009) and 4×10^6 km^2 (Yu et al. 2010). Tropical and southern peatlands likely cover an additional 0.4×10^6 km^2 (Yu et al. 2010) to 0.7×10^6 km^2 (Joosten 2010). Thus, peatlands likely represent approximately half of the total global wetland area of between $6–13 \times 10^6$ km^2 (see Mitsch and Gosselink 2007 and Melton et al. 2013 for range of wetland areas).

The carbon stored in these peatlands is related to their carbon content (mass of carbon per unit mass of soil), soil depth, and soil bulk density (mass of soil per unit volume). While it is conceptually and practically straightforward to measure these values at a single site,

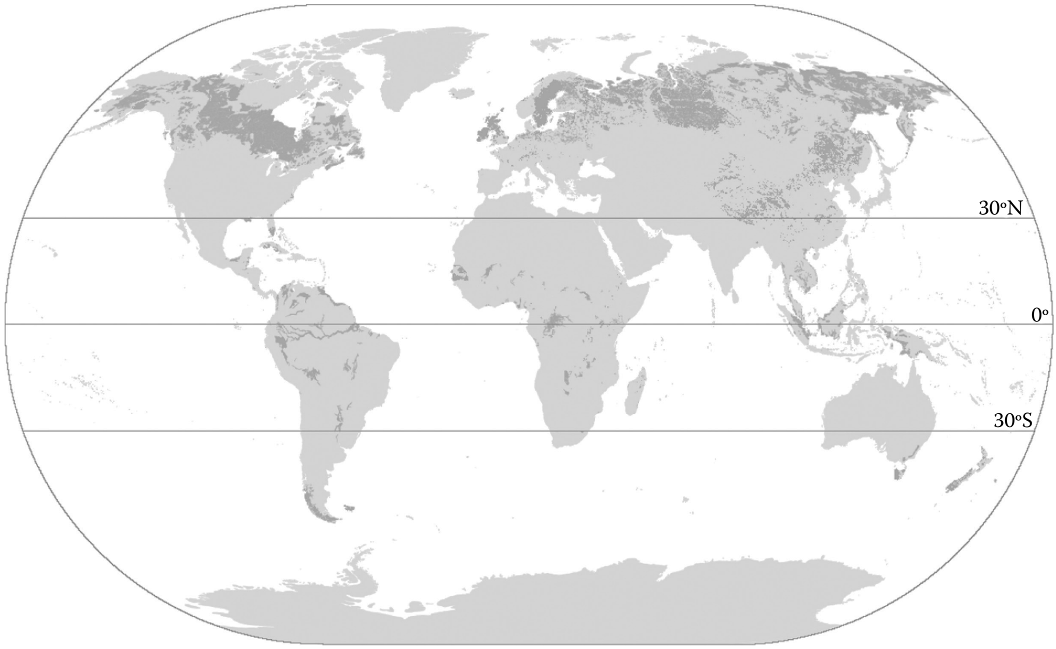


FIGURE 6.4
Map of global peatland area with peatlands indicated as shaded areas. (From Figure S1 in Yu, Z. et al. 2010. *Geophysical Research Letters* 37: L13402.)

working at the global scale introduces a number of assumptions. For example, many estimates of carbon density for peatlands are given to a depth of 1–2 m even though peatland soils can be much deeper. For example, Buffam et al. (2010) reported peat depths of >10 m in some Wisconsin peatlands. Bulk density and carbon content can also vary between different classes of peatlands and change with depth within a peatland ecosystem. In their estimates of North American peatland soil carbon stocks, Bridgham et al. (2006) used a carbon density of 1500 Mg C ha^{-1} for peatlands in the conterminous United States and Mexico (this value was an average of published values for many peatland types). Tarnocai et al. (2005) used values of 1441 and 1048 Mg C ha^{-1} for Canadian histosols and histels, respectively (these values were also used by Bridgham et al. 2006). Page et al. (2011) used a value of 1100 Mg C ha^{-1} for peats to a depth of 1.5 m and a value of 1466 Mg C ha^{-1} for peats to a depth of 2.3 m.

Using the carbon density approach has generated a range of values for the global pool of carbon in peat soils. Bridgham et al. (2006) used a value of 462 (Pg C) for permafrost and non-permafrost peatlands combined. This estimate was derived from the estimate of Maltby and Immirzi (1993), and represents a value in the middle of the 234–679 Pg carbon range reported from a number of published studies (see Bridgham et al. 2006 for details). Page et al. (2011) provided 610 Pg C as their best estimate for the global peat carbon pool.

While peatland soils dominate wetland soil carbon storage, non-peatland wetland soils also store carbon. Bridgham et al. (2006) estimated that despite a carbon density of only 199 Mg C ha^{-1} (a value at least five times lower than for peat soils; Batjes 1996), an additional 46 Pg C is stored in freshwater mineral wetlands as a result of their large global area.

Carbon storage in coastal wetland ecosystems including salt marshes, mangroves, and seagrass beds (the so-called blue carbon) has also received considerable attention recently (Chmura et al. 2003; Mcleod et al. 2011; Pendleton et al. 2012). Chmura et al. (2003) estimated a carbon storage of 430 Tg (0.43 Pg) in the surface 50 cm of global salt marsh ecosystems. This estimate was derived from an estimated salt marsh area of 2.2 Mha and an average carbon density of 195 Mg C ha^{-1} (assuming a 50 cm depth) for salt marsh soils (Chmura et al. 2003). Using a higher carbon density of 250 Mg C ha^{-1} (to a depth of 1 m) and an estimate of global salt marsh area of 5.1 Mha, Pendleton et al. (2012) calculated a higher estimate of 1.3 Pg C, reflecting the assumption of deeper soil depths, increased area, and a higher carbon density. Compared to salt marshes, mangroves are likely to have higher carbon densities, with recent estimates as high as 784 Mg C ha^{-1} for belowground soil carbon storage to an average depth of 2 m (Donato et al. 2011). When multiplied by recent estimates of mangrove area (Giri et al. 2011), this translates to a global carbon pool of 11 Pg C. This value is considerably higher than the estimate of 5 Pg in mangrove soils provided by Chmura et al. (2003), although these authors did stress that their estimate was based on a soil depth of 50 cm while acknowledging that many mangrove soils were likely much deeper. Seagrass ecosystems also represent a globally significant pool of soil carbon. Fourqurean et al. (2012) used a median estimate of 139.7 Mg C ha^{-1} as a carbon density of the surface 1 m of sediment to estimate a global pool of 4.2–8.4 Pg C in these ecosystems (global seagrass area is not well known, explaining the range in this estimate).

As a final caveat, it should be noted that the *size* of the carbon pool in a wetland (or type of wetland) does not reflect the *rate* of carbon accumulation in that wetland. For example, while peatlands dominate global wetland carbon pools, this carbon has been accumulating relatively slowly over thousands of years (Yu et al. 2010). This is in contrast to salt marsh SOM which contributes minimally to the global wetland SOM pool, but can accumulate quite rapidly as these systems accrete new soil to keep pace with sea level rise (Kirwan and Mudd 2012; Kirwan and Megonigal 2013).

Physical and Chemical Properties of Organic Soils

The physical properties of organic soils (bulk density, water holding capacity, and hydraulic conductivity) are driven in large part by their organic content and degree of decomposition. In addition to having higher organic matter content than their mineral counterparts, organic soils also generally have lower bulk density and higher water holding capacity. Hydraulic conductivity can vary widely in both organic and mineral soils, but the controls on soil hydrology differ between these soils types. Verry et al. (2011) provide an excellent synthesis of the physical properties of organic soils, and we summarize key points here.

Organic Content

As discussed above, organic carbon content is crucial to the definition of organic soils. Carbon is typically the dominant element of organic matter on a mass basis, and determination of organic carbon content is generally based on measurement of organic matter. Historically, organic matter content was measured by either the Walkley-Black (wet combustion) method or by ignition (dry combustion). However, the Walkley-Black method is no longer recommended as a result of the low recovery of soil carbon, and calculating organic matter by loss on ignition is the preferred method (Soil Survey Staff 2004, 2011).

In the loss on ignition method, a dry soil sample (oven-dried to constant mass at 110°C for mineral soils; a 60°C temperature is often used for organic soils) is placed in a cold muffle furnace which is heated to 400°C for 16 h. At this high temperature, organic matter is ignited and lost as CO_2. The difference in mass before and after ignition is the organic matter content of the soil (%SOM)

$$\%\text{SOM} = [(\text{OD}_\text{W} - \text{R}_\text{W})/\text{OD}_\text{W}] \times 100 \tag{6.1}$$

Where OD_W is the oven-dry soil weight and R_W is the residue weight after ignition. SOM content is alternatively expressed as loss on ignition (%LOI; Nelson and Sommers 1996). Organic carbon content is typically calculated based on the assumption that SOM contains about 58% organic carbon.

$$\%\text{Organic carbon} = \%\text{SOM} \times 0.58 \tag{6.2}$$

This equation is built on the van Bemmelen factor of 1.724, which is used to convert measured values of %organic carbon to %SOM (i.e., the reverse of Equation 6.2). Regional- and site-specific differences in plant community composition and decomposition, which together influence SOM chemistry, are likely to be important in the relationship between SOM and %organic carbon and the accuracy of these general relationships should be viewed with caution. As has been noted elsewhere, the assumption that 58% of SOM is organic carbon is, on average, too high (Nelson and Sommers 1996; Pribyl 2010). Pribyl (2010) suggests that a value of 50% organic carbon may be more accurate in most cases.

Organic carbon content can be measured directly using elemental analyzers following the treatment of soils with acid to ensure that all inorganic carbon has been removed. It is thus possible through derived equations to calculate %organic carbon from %SOM values directly. For example, Craft et al. (1991) derived the following quadratic equation for estuarine marsh soils:

$$\%\text{Organic carbon} = (\%\text{SOM} \times 0.40) + (\%\text{SOM}^2 \times 0.0025) \tag{6.3}$$

Degree of Decomposition

Saturated Histosols are defined in large part by the degree of decomposition, from slightly decomposed Fibrists to moderately decomposed Hemists to highly decomposed Saprists (Figure 6.3). The degree of decomposition (Table 6.2) is measured through analysis of fiber volume and pyrophosphate color (Soil Survey Staff 2004, 2011).

Fiber volume can be estimated qualitatively in the field as described in Chapter 1, or more quantitatively in the laboratory. Laboratory methods subject a soil sample to increasing physical stress and record the volume of fibrous material that remains on a 100-mesh (0.152 mm) sieve. Rubbed fiber volume is critical for defining suborders of Histosols (Soil Survey Staff 2010). To determine rubbed fiber volume in the laboratory, moist soil is packed into a half syringe of known volume (typically 5 or 2.5 mL). The soil is initially washed over a 100-mesh sieve using tap water with a flow rate of 40–60 mL s^{-1} until the water passing through the sieve is clear. Additional washing following mixing with an egg beater (after the first wash) and subsequently a blender (after the second wash) is required if >10% sapric material remains in the washed sample (determined by visual inspection).

When <10% sapric material remains after any wash, the residue is returned to the syringe to measure the unrubbed fiber volume. The resulting unrubbed fiber is transferred to a 100-mesh screen and rubbed between thumb and fingers under a stream of tap water with a flow rate of 30–40 mL s^{-1} until water passing through the sieve is clean. The remaining rubbed fibers are rolled between the thumb and fingers. The final residue is returned to the syringe to measure rubbed fiber volume (Soil Survey Staff 2004). The rubbed fiber content of >40% (by volume) is associated with fibric materials and rubbed fiber content of <17% (by volume) is associated with sapric material. Hemic materials are intermediate in rubbed fiber volume (Table 6.2; Soil Survey Staff 2010).

In pyrophosphate color analysis, a sample of moist soil is mixed and allowed to equilibrate with a saturated sodium pyrophosphate solution. The color of the resulting mixture correlates with the degree of decomposition. A strip of chromatographic paper is moistened with the solution and compared to the 7.5YR or 10YR Munsell color charts (Soil Survey Staff 2004). Light colors are associated with fibric materials, and dark colors are associated with sapric materials. Intermediate colors are associated with hemic materials (Table 6.2; Soil Survey Staff 2010).

A number of additional metrics are available to measure the degree of decomposition of organic soils. These include the ASTM's fiber content method (American Society for Testing Material 2008), which is similar to the method described above except that peat is initially soaked for 15 h in a dispersing agent (sodium hexametaphosphate) and fiber content is expressed on a dry-mass basis. Also included is the centrifugation method (Parent and Caron 2008), based on the USSR method, which measures the volume of peat passing through a sieve fitted over a centrifuge tube. The degree of decomposition is frequently measured in the field using the von Post method (Parent and Caron 2008), which identifies 10 humification degrees (H-values) defined by squeezing fresh soil in the palm of the hand and noting the color and consistency of the extruded water and residue. Details of the humification classes are described elsewhere (Malterer et al. 1992; Parent and Caron 2008; Verry et al. 2011), but lower values refer to less-decomposed SOM and higher values refer to more-decomposed SOM. As described in Malterer et al. (1992) and summarized in Verry et al. (2011), the von Post and USSR methods are capable of more accurately defining a larger number of decomposition classes. However, there are reasonably strong predictive relationships between these different metrics, demonstrating that a variety of methods are appropriate for determining the degree of decomposition of organic soils. Degree of decomposition can be estimated qualitatively into the field as described in Chapter 1.

Bulk Density

Bulk density is the mass of dry soil per unit bulk volume and reflects the porosity, compaction, and mineral content of a soil. In organic soils, a core of undisturbed soil is collected with a core sampler or a McCauley sampler of known volume and dried to a constant mass at 105°C (some methods use 110°C; Caron et al. 2008). Minimizing soil compaction, especially of surface soils, can be a challenge when working with organic soils. A number of comparative studies of organic soils suggest that bulk densities generally range from ~0.02 to 0.35 Mg m^{-3} (Bridgham et al. 1998; Verry et al. 2011). Bulk density is known to decrease with increasing organic content across soils ranging from mineral to organic (Ruehlmann and Körschens 2009) and within organic soils (Verry et al. 2011 and references cited therein). Within organic soils, bulk density generally increases with degree of decomposition, with higher bulk densities found at higher von

Post categories and in more sapric peats with lower fiber content (Verry et al. 2011 and references cited therein).

Water Content, Water Retention, and Hydraulic Conductivity

Soil water content is the water lost from the soil after drying to a constant mass at 105°C (Soil Science Society of America 2008). Water content can be expressed in a number of different ways in organic soils based upon measurements of wet soil (Wr), dry soil (Wd), and mass of water lost (Ww) (Ww = Wr – Wd). As described by Verry et al. (2011), water content (%) in mineral soils is generally expressed as the ratio of the mass of water lost to the oven-dry mass of soils (Ww/Wd × 100). Water content generally increases with an organic carbon content of soils and can be quite large in organic soils, ranging from 300% to 3000%. This makes comparisons to mineral soils (typically <100% water content) difficult. The water content of organic soils has also been expressed on the basis of bulk saturated mass (Ww/Wr × 100) or per unit volume of bulk soil (Verry et al. 2011). The water content of fully saturated peats can range from ~80% to nearly 100% on a per-volume basis, and generally decreases with increasing bulk density or degree of decomposition. Thus, sapric soils with higher bulk densities and lower fiber contents generally hold less water on a per volume basis (Verry et al. 2011).

The ability of organic soils to retain water is also related to bulk density and degree of decomposition. Water retention, expressed as % volume, is derived experimentally by subjecting soils to differential water tensions but is also related to drainage and field capacity of soils in natural settings. While sapric soils with high bulk densities and low fiber contents generally hold less water per unit volume than less-decomposed hemic and fibric soils, water is retained more effectively by sapric soils, suggesting that these soils drain less efficiently (Verry et al. 2011).

The hydraulic conductivity of soil describes the ease with which water can move through a soil and is important for regulating wetland hydrology (Chapter 3). Given their high water content and low bulk density, hydraulic conductivity is challenging to measure in organic soils (Caldwell et al. 2005). Hydraulic conductivity is generally lower in more-decomposed sapric soils with higher bulk density, fiber content, and von Post H-values (Verry et al. 2011). Recent work has also demonstrated that analytical measurements of peat chemistry (derived from ^{13}C-NMR data) better explain hydraulic conductivity in peats than simple measurements of bulk density (Grover and Baldock 2013). These authors acknowledge the ease of measuring bulk density but suggest that more sophisticated chemical analyses may be warranted when looking to obtain a more accurate prediction of peat hydraulic conductivity.

Ecology of Soil Organic Matter

As in all other ecosystems, the accumulation of organic matter in wetland soils is a result of an imbalance between organic matter gained and organic matter lost. The organic matter enters a wetland through endogenous net primary production (NPP) and transport of organic matter in both solid and dissolved forms. Losses of organic matter occur through decomposition and transport out of the wetland. This simple mass balance approach can be adapted to any element in organic matter (e.g., organic carbon or organic nitrogen);

however, the importance of the terms can vary for different elements. In this section, we initially focus on carbon cycling, with particular emphasis on the unique anaerobic decomposition dynamics that dominate in wetland soils. We end with a brief consideration of noncarbon elements in SOM.

Carbon Cycling

The ultimate source for carbon, and thus a majority of SOM, in wetland ecosystems, is NPP by photosynthesis, which reduces inorganic CO_2 into organic carbon. The amount of carbon that enters a wetland through NPP is highly variable, with a range of 10–4600 g C m^{-2} y^{-1} for aboveground NPP (Gopal and Masing 1990). Aboveground NPP in northern peatlands is also quite variable, but is generally <1000 g C m^{-2} y^{-1} (e.g., Szumigalski 1996; Thormann 1999; Vitt et al. 2001; Weishampel et al. 2009). In some wetlands, including the Florida Everglades, periphyton can contribute substantially to NPP. Ewe et al. (2006) monitored annual NPP of dominant primary producers across 17 sites within the Florida Coastal Everglades fresh-estuarine gradient, and demonstrated that while highly variable, periphyton productivity could dominate overall productivity, exceeding macrophyte productivity by an order of magnitude. These values do not include belowground NPP, which is rarely measured due in large part to the logistical challenges associated with measuring belowground processes in wetland soils (Iversen et al. 2012). It is generally recognized however that belowground NPP can be significant. For example, Weltzin et al. (2000) used root ingrowth cores to show that belowground NPP was between 26%–60% and 55%–86% of total aboveground NPP in bog and fen mesocosms, respectively.

Landscape position likely exerts the largest degree of control over the amount of organic matter that is transported to or from a wetland. Within Histosols, bog systems are dominated by autochthonous organic matter production whereas minerogenous fens may capture allochthonous organic matter or transport dissolved organic matter due to a larger degree of landscape connectivity. As mentioned above, despite low carbon density, sedimentation can be an important input of organic matter in nontidal freshwater wetlands (Bridgham et al. 2006) and can affect the ecology of many coastal wetlands (e.g., Neubauer et al. 2002; Neubauer 2008). Transport of dissolved carbon can also influence the carbon balance of many peatland soils (e.g., Pastor et al. 2003; Freeman et al. 2004; Yu 2012).

The decomposition of organic matter begins when above- or belowground material from primary producers senesces and is processed by a diverse soil community (Chapter 5). In many cases, this senescent material is recognizable as plant structures (e.g., leaves) and is initially broken down into smaller fragments through a series of biological and physical processes, including processing by heterotrophic invertebrates. This fragmentation increases the surface area for colonization by a diverse microbial community including bacteria, fungi, and an associated microbial food web of "predators," like amoebae, that consume these fungi and bacteria (e.g., Lin et al. 2012; Andersen et al. 2013; Jassey et al. 2013). The initial step in microbial decomposition involves the degradation of complex molecules into monomeric forms by hydrolytic enzymes. The main enzymes involved in cellulose degradation include exocellulase, endocellulase, and β-glucosidase (Sinsabaugh et al. 1991). Organic matter decomposition is augmented by enhanced lignin degradation through the production of phenol oxidase by microorganisms, including fungi (Sinsabaugh et al. 1991). These monomeric organic molecules are ultimately used as electron donors by heterotrophic microbes. Above the water table in wetlands, O_2 is available as a terminal electron acceptor and organic carbon is mineralized to CO_2 aerobically. Under O_2-rich aerobic conditions, it is possible for a single microorganism to completely decompose complex organic matter and decomposition proceeds at an accelerated rate, resulting in high turnover and production of CO_2.

Anaerobic Carbon Cycling

Below the water-table, O_2 is generally not available as a terminal electron acceptor and the complete mineralization of organic matter requires a complex community of microbes (Megonigal et al. 2004; Bridgham et al. 2013) ultimately generating CO_2 and CH_4 as respiratory by-products (Figure 6.5, Chapter 4). CH_4 is an important greenhouse gas with 45-times the sustained-flux global warming potential of CO_2 (over a 100-year timeframe; Neubauer and Megonigal 2015). Wetlands are the largest natural source of CH_4 to the atmosphere, and there is a long history of attempting to understand the role of wetlands in the global CH_4 cycle, especially in the context of climate change (Bridgham et al. 2013).

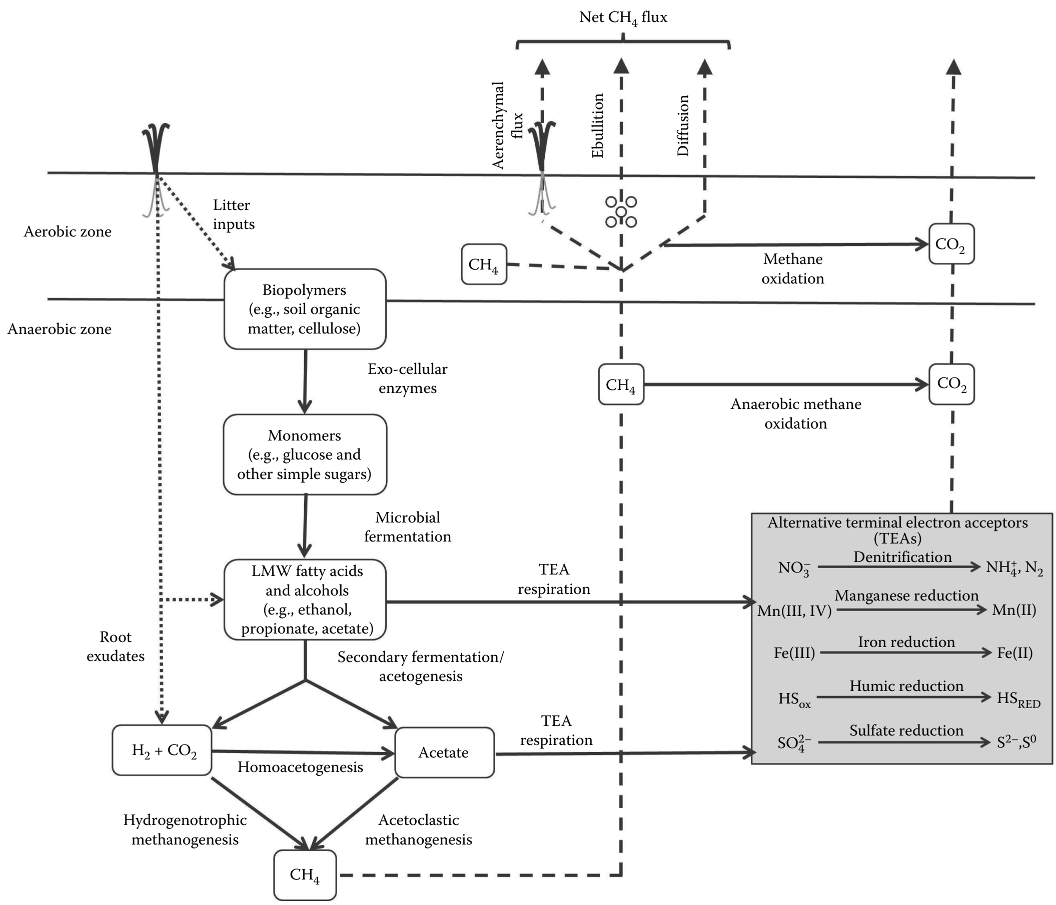


FIGURE 6.5
Decomposition in wetland ecosystems. Pools of carbon are shown in white boxes, and solid arrows show the progressive mineralization of these carbon pools by the identified microbial processes or groups. Dotted lines illustrate carbon inputs from the plant community. Dashed lines represent the flux of the gaseous end products of these processes (CH_4 and CO_2) to the atmosphere. (Modified from Bridgham, S. D. et al. 2013. *Global Change Biology* 19: 1325–1346, with permission.)

The initial steps of anaerobic microbial decomposition are similar to those in aerobic environments, namely microbes release extracellular enzymes in the soil environment to degrade complex organic matter into monomeric units. One fundamental difference, however, is the fact that the enzyme phenol oxidase appears to be highly sensitive to oxygen concentrations and may be inactive under anaerobic conditions. Phenol oxidase is a key enzyme that regulates ecosystem function through its fundamental role in lignin degradation (carbon and additional nutrient acquisition), soil humification, and dissolved organic carbon export (Sinsabaugh 2010). The inactivity of this key enzyme under anaerobic conditions can result in an accumulation of toxic phenolic compounds that inhibit other enzymes and subsequently slow SOM decomposition (e.g., Pind et al. 1994; Freeman et al. 2001; Fenner and Freeman 2011; Bragazza et al. 2012). Additional controls on phenol oxidase activity include moisture content, temperature, and pH, which can interact with oxygen availability in regulating enzyme activity (Sinsabaugh 2010).

Organic molecules released through extracellular enzyme activity are transported across microbial cell membranes and are taken up by microbes prior to being further processed by fermentation. Fermentation involves the transfer of electrons within an organic molecule and results in the splitting of larger organic compounds into a series of progressively smaller fermentation products including low molecular weight organic acids, alcohols, di-hydrogen (H_2), and CO_2. Among the most important fermentation products in wetland soils are H_2, CO_2, and acetate (Conrad 1999 and references cited therein).

Reduced fermentation products (i.e., H_2, acetate, and other organic acids) serve as electron donors and are oxidized to CO_2 and/or H_2O by heterotrophic microbial respiration using a variety of alternative terminal electron acceptors (TEAs) in place of O_2 (Chapter 4). The microbial community generally uses TEAs preferentially in the following order of decreasing thermodynamic efficiency: NO_3^- (denitrification); Mn(IV, III) (manganese reduction); Fe(III) (iron reduction); and SO_4^{2-} (sulfate reduction). From a thermodynamic perspective, the reduction of these TEAs is competitively favorable to the production of CH_4 by methanogenesis (discussed further below) and CH_4 production is limited in the presence of these TEAs.

The relative importance of these various microbial processes differs based on the availability of TEAs in different wetland ecosystems. For example, while iron reduction can dominate anaerobic respiration in mineral wetland soils (e.g., Lovley and Phillips 1986; Roden and Wetzel 2003), it plays a less important role in organic wetland soils, which generally lack iron. Sulfate reduction generally dominates in saline and brackish wetlands due to a continuous supply of SO_4^{2-} from tidal exchange, and helps explain the low CH_4 production in these environments (Bartlett et al. 1987; Poffenbarger et al. 2011). Somewhat surprisingly, despite low SO_4^{2-} concentrations, microbial sulfate reduction has also been shown to play an important role in anaerobic carbon cycling in many freshwater peatland ecosystems (Vile et al. 2003a,b) and sulfate deposition from industrial activities has been shown to lower CH_4 emissions at the ecosystem scale (Gauci et al. 2004). More recently, it has become apparent that organic matter (i.e., humic substances) can also serve as an alternative TEA for microbial respiration and may be linked to iron and sulfur cycling (Lovley et al. 1996; Heitmann and Blodau 2006; Heitmann et al. 2007; Bridgham et al. 2013; Martinez et al. 2013; and Klüpfel et al. 2014). The reduction of solid-phase SOM has been demonstrated to inhibit CH_4 production in wetland soils and can dominate anaerobic decomposition in these systems (Keller et al. 2009; Keller and Takagi 2013).

After more favorable inorganic (NO_3^-, Mn(IV,III), Fe(III), and SO_4^{2-}) and organic TEAs have been exhausted, CH_4 is formed by methanogenic archaebacteria. In wetland ecosystems, CH_4 is produced predominately through the acetoclastic and hydrogenotrophic pathways. Acetoclastic methanogens produce CH_4 and CO_2 from the fermentation of acetate. Hydrogenotrophic methanogens produce CH_4 through the reduction of CO_2 using H_2 as an electron donor (Bridgham et al. 2013). Methane can also be produced using a variety of other substrates (e.g., carbon monoxide, some alcohols, formate, and methylated compounds such as trimethylamine, dimethyl sulfate, and methanol) (Zinder 1993). These pathways are typically assumed to play a minor role in CH_4 cycling in natural ecosystems, although there is evidence that they can dominate CH_4 production in hypersaline environments (e.g., Potter et al. 2009; Kelley et al. 2012). In addition to competition with other TEA-reducing processes, methanogens can be influenced by the activity of homoacetogens, which produce acetate through the reduction of CO_2 (with H_2) rather than through the fermentation of more complex organic substrates (Figure 6.5). Thus, homoacetogenesis directly competes for substrates with hydrogenotrophic methanogens but produces acetate, the key substrate for acetoclastic methanogens. Recent evidence suggests that homoacetogenesis may play an important role in anaerobic carbon cycling in organic soils (Drake et al. 2009; Ye et al. 2014).

Once CH_4 is produced by methanogens, it can leave a wetland through diffusion, ebullition, and/or plant-mediated transport (Figure 6.5). The relative importance of these pathways plays an important role in controlling net CH_4 emissions to the atmosphere from a wetland. Perhaps most importantly, CH_4 leaving through the diffusive pathways is subject to consumption by chemoautotrophic methanotrophic bacteria that oxidize CH_4 to CO_2 (Hanson and Hanson 1996) and can consume a significant fraction of gross CH_4 production (see references in Megonigal et al. 2004; Bridgham et al. 2013). The emission of CH_4 via plant-mediated transport (i.e., through plant aerenchyma) or via ebullition generally allows CH_4 to bypass zones of methanotrophy with the end result that a larger fraction of gross CH_4 production is emitted to the atmosphere. Consumption of CH_4 by methanotrophs has historically been thought to be limited to aerobic regions of wetland environments where oxygen is available as a TEA, primarily above the water-table level and in the oxic rhizosphere of wetland plant communities. There is mounting evidence, however, that anaerobic CH_4 consumption is possible or even widespread in freshwater peatlands. The anaerobic oxidation of CH_4 would require the use of alternative TEAs; moreover, it has been suggested that SOM may serve in this role in some wetland environments (Smemo and Yavitt 2011; Gupta et al. 2013).

Beyond Carbon: An Example of Nitrogen in Soil Organic Matter

While the previous summary has focused on tracking the flow of carbon through wetland ecosystems, understanding the cycling of other SOM-associated macronutrients, such as nitrogen and phosphorus, is crucial to microorganisms and plants. We use nitrogen as an example of the important role of SOM cycling for noncarbon elements in wetland soils. Nitrogen can be incorporated into the soil via atmospheric deposition and through nitrogen fixation (dinitrogen reduction to NH_3) carried out by free-living and root-associated microorganisms. However, the mineralization (converting from organic into mineral forms) and immobilization (taking up inorganic nutrients to build biomass) associated with microbial processing of SOM is an important component of the nitrogen cycle in many wetlands (Chapter 4).

Research in marine systems suggests a 106:16:1 (carbon:nitrogen:phosphorus) ratio in planktonic marine organisms and the surrounding ocean, a ratio which has been applied to multiple ecosystems as a proxy for nutrient limitation (Redfield 1958). Recent studies determined stoichiometric C:N:P ratios of 186:13:1 in soil and 60:7:1 in soil microbial biomass

(Cleveland and Liptzin 2007). Within the soil, microorganisms must maintain an internal C:N ratio of ~8:1 (Cleveland and Liptzin 2007; Brady and Weil 2008). Approximately two-third of the carbon metabolized by microorganisms is respired as CO_2 thus, on average, microbes require 1 gram of nitrogen for every 25 grams of carbon consumed from SOM (Brady and Weil 2008). Throughout the process of organic matter decomposition, the C:N ratio often changes, typically declining over time (Reddy and DeLaune 2008). At higher C:N ratios in SOM (>25:1), microbes typically immobilize nitrogen from the environment. However, declining C:N ratios within SOM (<25:1) shift microbes from net immobilization to net mineralization of nitrogen. Thus, a decreasing C:N ratio within the organic matter can be an indicator of inorganic nitrogen release (mineralization) within the soil (Reddy and DeLaune 2008).

Human Use of Wetland Soil Organic Matter

Wetland SOM has long been considered a resource due to its high fertility and energy content, and, as such, human use has had an impact on SOM at the global scale. A number of recent reviews explore these impacts (Minkkinen et al. 2008; Oleszczuk et al. 2008; Joosten 2009; Laine et al. 2009). Here, we summarize key themes and illustrate potential impacts with selected case studies.

Agriculture remains a dominant human impact on northern peatlands worldwide (Oleszczuk et al. 2008; Laine et al. 2009), with wetland loss ascribed to agricultural activities estimated up to 18% in North America, 25% in China, and 15% in Europe (without Russia) (Baird et al. 2009 and references therein). Forestry is another major anthropogenic driver of wetland loss on a global scale (Baird et al. 2009) and large areas of peatlands have been drained for forestry, especially in northern Europe and Russia (Minkkinen et al. 2008; Laine et al. 2009). Peat extraction for fuel and horticultural purposes are additional unique uses of organic wetland soils but play a much smaller role than agricultural and forestry activities (Laine et al. 2009).

Common to agricultural, forestry, and peat extraction for fuel and horticultural use is a lowering of the existing water-table level with important consequences for SOM. Lowering of the water table can introduce oxygen to previously anaerobic peat, leading to increased oxidative losses of SOM as CO_2 while concomitantly reducing CH_4 emissions (Joosten 2009; Laine et al. 2009). This loss of organic matter following drainage, and to a lesser extent, consolidation of peat (Schothorst 1977; Hooijer et al. 2012), results in soil subsidence (Ewing and Vepraskas 2006). One of the best-studied examples of soil subsidence is the Florida Everglades (Florida, USA). Landscape drainage for agricultural use of the Everglades began in the 1880s (McVoy et al. 2011) resulting in a lowered water table and subsequent exposure of organic soils to aerobic conditions. Oxidation of organic matter has been observed with an estimated average subsidence rate of 1.6 m over an 88-year period. This subsidence rate equates to a 4×10^{12} m^3 loss of peat and an additional flux of 4.9×10^8 metric tons of CO_2 to the atmosphere (Aich et al. 2013).

Soil subsidence is also of concern in the Sacramento-San Joaquin Delta (California, USA), where subsidence threatens the water supply and infrastructure of more than 23 million California residents (Deverel and Leighton 2010). Drainage and cultivation of surrounding soils has resulted in subsidence ranging from 1 to more than 8 m below sea level since 1850, which has resulted in an estimated loss of 2×10^{12} m^3 of sediment marsh (Deverel and Leighton 2010, and references therein). Using elevation data, a median subsidence rate of 1.6–2.6 cm y^{-1} was estimated for the entire delta ($n = 2570$; Deverel and Leighton 2010).

Another form of disturbance (natural and anthropogenic) which can affect SOM storage is fire. For example, carbon loss from combustion of western Canadian peatlands averages 4.7 ± 0.6 Tg C y^{-1}, suggesting that fire can convert peatland landscapes, which are typically a carbon sink, into a substantial carbon source (Turetsky et al. 2002). This loss of carbon is equivalent to an estimated loss of 1470 km^2 annually of peatlands from this region (Turetsky et al. 2002). While important, the effect of this landscape disturbance is highly variable and controlled largely by soil conditions as well as fire characteristics such as fire type and intensity (González-Pérez et al. 2004).

The moisture content of organic peat soils is an important mediator of the response of SOM to fire. In a laboratory-based ignition experiment, Rein et al. (2008) adjusted the moisture content (from 85% to 160%) of peat monoliths from Edinburgh, Scotland to demonstrate that peat with moisture content above 135% was unable to ignite. When coupling these data with mass loss, more than 90% of the peat was oxidized in monoliths with a moisture content below 115% (Bodelier et al. 2006) which has implications for estimating fire-mediated carbon loss from boreal peatlands. The importance of soil moisture at the field scale was highlighted in work by Turetsky et al. (2011) who established replicate burned and unburned (control) plots in nondrained and experimentally drained (water table initially lowered by 25 cm) fen soils in western Canada. Carbon was lost from both drained and nondrained plots; however, carbon loss was more than eight fold greater from drained soils (16.8 and 2.0 kg C m^{-2}, respectively).

Fire intensity and duration also have large-scale implications for SOM storage and subsequent carbon loss. While flaming combustion can be shortlived at a given location, smoldering (flameless) combustion can continue for days, months, or even years in peatlands (Watts 2013, and references therein), which can strongly affect organic matter storage, greenhouse gas production, and ecosystem function. A field-based study of the Scottish Highlands investigated deep peat oxidation following smoldering combustion. Through quantification of total fuel (litter, duff, and peat), oxidative loss of organic matter was 773 ± 120 Mg, including 396 ± 63 Mg C (96 ± 15 t C ha^{-1}) (Davies et al. 2013) highlighting the large effect smoldering combustion can have on organic carbon loss across these ecosystems. With fire frequency expected to increase with climate change (Westerling et al. 2006; Krawchuk et al. 2009), understanding the interactive effects of altered hydrology and fire on organic matter loss and greenhouse gas emissions from peatlands is crucial.

Conclusion

The unique ecology of wetland ecosystems, and, in particular, the low-oxygen environments associated with flooded or saturated soils has allowed wetlands to accumulate globally significant amounts of organic matter in their soils. Perhaps not surprisingly, the majority of this wetland SOM is stored in peatland ecosystems, which are explicitly defined by highly organic soils. While all peatlands have organic soils by definition, there are dramatic differences in the characteristics of SOM between different types of peatlands, with important implications for their hydrology and ecology. Recent work has called into question a number of long-held assumptions about SOM. In particular, the structure and longevity of humic substances have been challenged, although these changing views have not yet been fully explored in wetland environments. Mounting evidence for the role of SOM as a terminal electron acceptor is also changing our understanding of anaerobic wetland decomposition. Ongoing human use of wetland SOM generates significant impacts and will likely become more important and complex in the face of ongoing global change.

References

Aich, S., C. W. McVoy, T. W. Dreschel, and Santamaria, F. 2013. Estimating soil subsidence and carbon loss in the Everglades Agricultural Area, Florida using geospatial techniques. *Agriculture, Ecosystems & Environment* 171: 124–133.

American Society for Testing Material. 2008. Standard test method for laboratory determination of the fiber content of peat sample y dry mass. *ASTM Standard D1997-91 2008.*

Amundson, R. 2001. The carbon budget in soils. *Annual Review of Earth and Planetary Sciences* 29: 535–562.

Andersen, R., S. J. Chapman, and R. R. E. Artz. 2013. Microbial communities in natural and disturbed peatlands: A review. *Soil Biology & Biochemistry* 57: 979–994.

Baird, A. J., L. R. Belyea, X. Comas, A. Reeve, and L. D. Slater. 2009. *Carbon Cycling in Northern Peatlands*. American Geophysical Union, Washington, DC, USA.

Bartlett, K. B., D. S. Bartlett, R. C. Harriss, and D. I. Sebacher. 1987. Methane emissions along a salt marsh salinity gradient. *Biogeochemistry* 4: 183–202.

Bartlett, K. B. and R. C. Harriss. 1993. Review and assessment of methane emissions from wetlands. *Chemosphere* 26: 261–320.

Batjes, N. H. 1996. Total carbon and nitrogen in the soils of the world. *European Journal of Soil Science* 47: 151–163.

Bodelier, P. L. E. et al. 2006. Ecological aspects of microbes and microbial communities inhabiting the rhizosphere of wetland plants. In J. T. A. Verhoeven, B. Beltman, R. Bobbink, and D. F. Whigham (Eds.) *Ecological Studies, Vol. 190. Wetlands and Natural Resource Management.* Springer_Verlag, Berlin, Heidelberg.

Brady, N. C. and R. R. Weil. 2008. The Nature and Property of Soils, 14th edition. Person Education, Inc., Upper Saddle River, New Jersey.

Bragazza, L., J. Parisod, A. Butler, and R. D. Bardgett. 2012. Biogeochemical plant-soil microbe feedback in response to climate warming in peatlands. *Nature Climate Change* 3: 273–277.

Bridgham, S. D., H. Cadillo-Quiro, J. K. Keller, and Q. Zhuang. 2013. Methane emissions from wetlands: Biogeochemical, microbial, and modeling perspectives from local to global scales. *Global Change Biology* 19: 1325–1346.

Bridgham, S. D., J. P. Megonigal, J. K. Keller, N. B. Bliss, and C. Trettin. 2006. The carbon balance of North American wetlands. *Wetlands* 26: 889–916.

Bridgham, S. D., J. Pastor, J. Janssens, C. Chapin, and T. Malterer. 1996. Multiple limiting gradients in peatlands: A call for a new paradigm. *Wetlands* 16: 45–65.

Bridgham, S. D., K. Updegraff, and J. Pastor. 1998. Carbon, nitrogen, and phosphorus mineralization in northern wetlands. *Ecology* 79: 1545–1561.

Buffam, I., S. R. Carpenter, W. Yeck, P. C. Hanson, and M. G. Turner. 2010. Filling holes in regional carbon budgets: Predicting peat depth in a north temperate lake district. *Journal of Geophysical Research* 115: G01005.

Caldwell, P.V., A. A. Adams, C. P. Niewoehner, M. J. Vepraskas, and J. D. Gregory. 2005. Sampling device to extract intact cores in saturated organic soils. *Soil Sci. Soc. Am. J.* 69: 2071–2075.

Caron, J., D. E. Elrich, J. C. Michel, and R. Naasz. 2008. Physical properties of organic soils and growing media: Water and air storage and flow dynamics. In M. R. Carter, and E. G. Gregorich (Eds.) *Soil Sampling and Methods of Analysis.* CRC Press, Taylor & Francis Group, Boca Raton, FL, pp. 885–912.

Chmura, G. L., S. C. Anisfel, D. R. Cahoon, and J. C. Lynch. 2003. Global carbon sequestration in tidal, saline wetland soils. *Global Biogeochemical Cycles* 17: 1111, doi: 10.1029/2002GB001917, 4.

Cleveland, C. C. and D. Liptzin. 2007. C: N: P stoichiometry in soil: Is there a “Redfield ratio” for the microbial biomass? *Biogeochemistry* 85: 235–252.

Conrad, R. 1999. Contribution of hydrogen to methane production and control of hydrogen concentrations in methanogenic soils and sediments. *FEMS Microbiology Ecology* 28: 193–202.

Craft, C. B., E. D. Seneca, and S. W. Broome. 1991. Loss on ignition and Kjeldahl digestion for estimating organic carbon and total nitrogen in estuarine marsh soils: Calibration with dry combustion. *Estuaries* 2: 175–179.

Davies, G. M., A. Gray, G. Rein, and C. J. Legg. 2013. Peat consumption and carbon loss due to smouldering wildfire in a temperate peatland. *Forest Ecology and Management* 308: 169–177.

Deverel, S. J. and D. A. Leighton. 2010. Historic, recent, and future subsidence, Sacramento-San Joaquin Delta, California, USA. *San Francisco Estuary and Watershed Science* 8: 1–23.

Donato, D. C., J. B. Kauffman, D. Murdiyarso, S. Kurnianto, M. Stidham, and M. Kannine. 2011. Mangroves amont the most carbon-rich forests in the tropics. *Nature Geoscience* 4: 293–297.

Drake, H. L., M. A. Horn, and P. K. Wüst. 2009. Intermediary ecosystem metabolism as a main driver of methanogenesis in acidic wetland soil. *Environmental Microbiology Reports* 1: 307–318.

Duddleston, K. N., M. A. Kinney, R. P. Kiene, and M. E. Hines. 2002. Anaerobic microbial biogeochemistry in a northern bog: Acetate as a dominant metabolic end product. *Global Biogeochemical Cycles* 16: 1063.

Dungait, J. A. J., D. W. Hopkins, A. S. Gregory, and A. P. Whitmore. 2012. Soil organic matter turnover is governed by accessibility not recalcitrance. *Global Change Biology* 18: 1781–1796.

Ewe, S. M., E. E. Gaiser, D. L. Childers, D. Iwaniec, V. H. Rivera-Monroy, and R. R. Twilley. 2006. Spatial and temporal patterns of aboveground net primary productivity (ANPP) along two freshwater-estuarine transects in the Florida Coastal Everglades. *Hydrobiologia* 569: 459–474.

Ewing, J.M. and M. J. Vepraskas. 2006. Estimating primary and secondary subsidence in an organic soil 15, 20, and 30 years after drainage. *Wetlands* 26: 119–130.

Fenner, N. and C. Freeman. 2011. Drought-induced carbon loss in peatlands. *Nature Geoscience* 4: 895–900.

Fourqurean, J. W. et al. 2012. Seagrass ecosystems as a globally significant carbon stock. *Nature Geoscience* 5: 505–509.

Freeman, C. et al. 2004. Export of dissolved organic carbon from peatlands under elevated carbon dioxide levels. *Nature* 430: 195–198.

Freeman, C., N. Ostle, and H. Kang. 2001. An enzymic ‘latch’ on a global carbon store—A shortage of oxygen locks up carbon in peatlands by restraining a single enzyme. *Nature* 409: 149–149.

Gauci, V. et al. 2004. Sulfur pollution suppression of the wetland methane source in the 20th and 21st centuries. *Proceedings of the National Academy of Science* 101: 12583–12587.

Giri, C. et al. 2011. Status and distribution of mangrove forests of the world using earth observation satellite data. *Global Ecology and Biogeography* 20: 154–159.

González-Pérez, J. A., F. J. González-Vila, G. Almendros, and H. Knicker. 2004. The effect of fire on soil organic matter—A review. *Environment International* 30: 855–870.

Gopal, B. and V. Masing. 1990. Biology and ecology. In B. C. Pattan (Ed.) *Wetlands and Sahllow Continental Water Bodies.* Vol. 1. SPB Academic Publisching, The Netherlands, pp. 91–239.

Grover, S. P. P. and J. A. Baldock. 2013. The link between peat hydrology and decomposition: Beyond von Post. *Journal of Hydrology* 479: 130–138.

Gupta, V., K. A. Smemo, J. B. Yavitt, D. Fowle, B. Branfireun, and N. Basiliko. 2013. Stable isotopes reveal widespread anaerobic methane oxidation across latitude and peatland type. *Environmental Science & Technology* 47: 8273–8279.

Hanson, R. S. and T. E. Hanson. 1996. Methanotrophic bacteria. *Microbiological Reviews* 60: 439–471.

Heitmann, T. and C. Blodau. 2006. Oxidation and incorporation of hydrogen sulfide by dissolved organic matter. *Chemical Geology* 235: 12–20.

Heitmann, T., T. Godhammer, J. Beer, and C. Blodau. 2007. Electron transfer of dissolved organic matter and its potential significance for anaerobic respiration in a northern bog. *Global Change Biology* 13: 1771–1785.

Hooijer, A. et al. 2012. Subsidence and carbon loss in drained tropical peatlands. *Biogeosciences* 9: 1053–1071.

Iversen, C. M. et al. 2012. Advancing the use of minirhizotrons in wetlands. *Plant and Soil* 352: 23–39.

Jassey, V. E. J. et al. 2013. Above- and belowground linkages in *Sphagnum* peatland: Climate warming affects plant-microbial interactions. *Global Change Biology* 19: 811–823.

Jobbágy, E. G. and R. B. Jackson. 2000. The vertical distribution of soil organic carbon and its relation to climate and vegetation. *Ecological Applications* 10: 423–436.

Jones, D. L., C. Nguyen, and R. D. Finlay. 2009. Carbon flow in the rhizosphere: Carbon trading at the soil-root interface. *Plant and Soil* 321: 5–33.

Joosten, H. 2009. Human impacts: Farming, fire, forestry and fuel. In E. Maltby, and T. Barker (Eds.) *The Wetlands Handbook*. Wiley-Blackwell, Oxford, UK, pp. 689–718.

Joosten, H. 2010. The global peatland CO_2 picture: Peatland status and drainage related emissions in all countries of the world. Report for Wetlands International, Waegningen, The Netherlands.

Kelleher, B. P. and A. J. Simpson. 2006. Humic substances in soils: Are they really chemically distinct? *Environmental Science & Technology* 40: 4605–4611.

Keller, J. K. and K. K. Takagi. 2013. Solid-phase organic matter reduction regulates anaerobic decomposition in bog soil. *Ecosphere* 4: 54.

Keller, J. K., P. B. Weisenhorn, and J. P. Megonigal. 2009. Humic acids as electron acceptors in wetland decomposition. *Soil Biology & Biochemistry* 41: 1518–1522.

Kelley, C. A., J. A. Poole, A. M. Tazaz, J. P. Chanton, and B. M. Bebout. 2012. Substrate limitation for methanogenesis in hypersaline environments. *Astrobiology* 12: 89–97.

Kirwan, M. L. and J. P. Megonigal. 2013. Tidal wetland stability in the face of human impacts and sea-level rise. *Nature* 504: 53–60.

Kirwan, M. L. and S. M. Mudd. 2012. Response of salt-marsh carbon accumulation to climate change. *Nature* 489: 550–553.

Kleber, M. 2010. What is recalcitrant soil organic matter? *Environmental Chemistry* 7: 320–332.

Klüpfel, L., A. Piepenbrock, A. Kappler, and M. Sander. 2014. Humic substances as fully regenerable electron acceptors in recurrently anoxic environments. *Nature Geoscience* 7: 195–200.

Kolditz, K., O. Dellwig, J. Barkowski, M. Beck, H. Freund, and H.-J. Brumsack. 2009. Effects of de-embankment on pore water geochemistry of salt marsh sediments. *Journal of Coastal Research* 25: 1222–1225.

Krawchuk, M. A., S. G. Cumming, and M. D. Flannigan. 2009. Predicted changes in fire weather suggest increases in lightning fire initiation and future area burned in the mixedwood boreal forest. *Climatic change* 92: 83–97.

Laine, J., K. Minkkinen, and C. Trettin. 2009. Direct human impacts on the peatland carbon sink. Geophyisical Monograph Series 184. In A. J. Baird, L. R. Belyea, X. Comas, A. S. Reeve, and L. D. Slater (Eds.) *Carbon Cycling in Northern Peatlands*. American Geophysical Union, Washington, DC, pp. 71–78.

Lehmann, J., D. Solomon, J. Kinyangi, L. Dathe, S. Wirick, and C. Jocobsen. 2008. Spatial complexity of soil organic matter forms at nanometre scales. *Nature Geoscience* 1: 238–242.

Lin, X. et al. 2012. Microbial community structure and activity linked to contrasting biogeochemical gradients in bog and fen environments of the Glacial Lake Agassiz Peatland. *Applied and Environmental Microbiology* 78: 7023–7031.

Lipson, D. A., D. Zona, T. K. Raab, F. Bozzolo, M. Mauritz, and W. C. Oechel. 2012. Water-table height and microtopography control biogeochemical cycling in an Arctic coastal tundra ecosystem. *Biogeosciences* 9: 577–591.

Lovley, D. R., J. D. Coates, E. L. Blunt-Harris, E. J. P. Phillips, and J. C. Woodward. 1996. Humic substances as electron acceptors for microbial respiration. *Nature* 382: 445–448.

Lovley, D. R. and E. J. P. Phillips. 1986. Organic matter mineralization with reduction of ferric iron in anaerobic sediments. *Applied and Environmental Microbiology* 51: 683–689.

Lynch, J. M. and J. M. Whipps. 1990. Substrate flow in the rhizosphere. *Plant and Soil* 129: 1–10.

Maltby, E. and P. Immirzi. 1993. Carbon dynamics in peatlands and other and other wetland soils, regional and global perspectives. *Chemosphere* 27: 999–1023.

Malterer, T. J., E. S. Verry, and J. Erjavec. 1992. Fiber content and degree of decomposition in peats: A review of national methods. *Soil Science Society of America Journal* 56: 1200–1211.

Martinez, C. M., L. H. Alvarez, L. B. Celis, and F. J. Cervantes. 2013. Humus-reducing microorganisms and their valuable contribution in environmental processes. *Applied Microbilogy and Biotechnology* 97: 10293–10308.

Mcleod, E. et al. 2011. A blueprint for blue carbon: Toward an improved understanding of the role of vegetated coastal habitats in sequestering CO_2. *Frontiers in Ecology and the Environment* 9: 552–560.

McVoy, C. W., W. P. Said, J. Obeysekera, J. A. VanArman, and T. W. Dreschel. 2011. *Landscapes and Hydrology of the Predrainage Everglades*. University Press of Florida, Gainesville, FL, USA.

Megonigal, J. P., M. E. Hines, and P. T. Visscher, 2004. Anaerobic metabolism: Linkages to trace gases and aerobic processes. In W. H. Schlesinger (Ed.) *Biogeochemistry*. Elsevier-Pergamon, Oxford, UK, pp. 317–424.

Melillo, J. M., J. D. Aber, A. E. Linkins, A. Ricca, B. Fry, and K. J. Nadelhoffer. 1989. Carbon and nitrogen dynamics along the decay continuum: Plant litter to soil oganic matter. *Plant and Soil* 115: 189–198.

Mellegård, H., T. Stalheim, V. Hormazaba, P. E. Granum, and S. P. Hardy. 2009. Antibacterial activity of sphagnum acid and other phenolic compounds found in Sphagnum papillosum against food-borne bacteria. *Letters in applied microbiology* 49: 85–90.

Melton, J. R. et al. 2013. Present state of global wetland extent and wetland methane modelling: Conclusions from a model inter-comparison project (WETCHIMP). *Biogeosciences* 10: 753–758.

Miltner, A., P. Bombach, B. Schmidt-Brücken, and M. Kästner. 2012. SOM genesis: Microbial biomass as a significant source. *Biogeochemistry* 111: 41–55.

Minkkinen, K., K. A. Byrny, and C. Trettin. 2008. Climate impacts of peatland forestry. In M. Strack (Ed.) *Peatlands and Climate Change*. International Peat Society, Jyväskylä, Finland, pp. 97–122.

Mitsch, W. J. and J. G. Gosselink. 2007. *Wetlands*. John Wiley & Sons, Inc., New York, NY.

Näsholm, T., K. Kielland, and U. Ganeteg. 2009. Uptake of organic nitrogen by plants. *New Phytologist* 182: 31–48.

Natural Resources Conservation Service. 2010. *Field Indicators of Hydric Soils in the United Stats, Version 7.0*. Natural Resources Conservation Service, Washington, DC.

Nelson, D. W. and L. E. Sommers. 1996. Total carbon, organic carbon, and organic matter. In D. L. Sparks (Ed.) *Methods of Soil Analysis: Part 3. Chemical Methods*. Soil Science Society of America, Inc, Madison, Wisconsin, pp. 961–1010.

Neubauer, S. C. 2008. Contributions of mineral and organic components to tidal freshwater marsh accretion. *Estuarine, Coastal and Shelf Science* 78: 78–88.

Neubauer, S. C., Megonigal, J. P. 2015. Moving beyond global warming potential to quantify the climatic role of ecosystems. *Ecosystems*, doi: 10.1007/s10021-015-9879-4

Neubauer, S. C., I. C. Anderson, J. A. Constantine, and S. A. Kuehl. 2002. Sediment deposition and accretion in a mid-Atlantic (U.S.A.) tidal freshwater marsh. *Estuarine, Coastal, and Shelf Science* 54: 713–727.

Oleszczuk, O., K. Regina, H. Szajdak, H. Höper, and V. Maryganova. 2008. Impacts of agricultural utilization of peat soils on the greenhouse gas balance. In M. Strack (Ed.) *Peatlands and Climate Change*. International Peat Society, Jyväskylä, Finland, pp. 70–97.

Page, S. E., J. O. Rieley, and C. J. Banks. 2011. Global and regional importance of the tropical peatland carbon pool. *Global Change Biology* 17: 798–818.

Parent, L. E. and J. Caron. 2008. Physical properties of organic soils and growing media: Particle size and degree of decomposition. In M. R. Carter, and E. G. Gregorich (Eds.) *Soil Sampling and Methods of Analysis*. CRC Press, Taylor & Francis Group, Boca Raton, FL, pp. 871–883.

Pastor, J. et al. 2003. Global warming and the export of dissolved organic carbon from boreal peatlands. *Oikos* 100: 380–386.

Pendleton, L. et al. 2012. Estimating global "blue carbon" emissions from conservation and degradation of vegetated coastal ecosystems. *Plos One* 7: e43542.

Pind, A., C. Freeman, and M. A. Lock. 1994. Enzymic degradation of phenolc materials in peatlands—Measurement of phenol oxidase activity. *Plant and Soil* 159: 227–231.

Poffenbarger, H. J., B. A. Needelman, and J. P. Megonigal. 2011. Salinity influence on methane emissions from tidal marshes. *Wetlands* 31: 831–842.

Potter, E. G., B. M. Bebout, and C. A. Kelley. 2009. Isotopic composition of methane and inferred methanogenic substrates along a salinity gradient in a hypersaline microbial mat system. *Astrobiology* 9: 383–390.

Pribyl, D. W. 2010. A critical review of the conventional SOC to SOM conversion factor. *Geoderma* 156: 75–83.

Reddy, K. R. and R. D. DeLaune. 2008. *Biogeochemistry of Wetlands*. CRC Press, Taylor & Francis Group, LLC, Boca Raton, FL.

Redfield, A. C. 1958. The biological control of chemical factors in the environment. *Am. Sci.* 46: 205–221.

Rein, G., N. Cleaver, C. Ashton, P. Pironi, and J. L. Torero. 2008. The severity of smouldering peat fires and damage to the forest soil. *Catena* 74: 304–309.

Roden, E. E. and R. G. Wetzel. 2003. Competition between Fe(III)-reducing and methanogenic bacteria for acetate in iron-rich freshwater sediments. *Microbial Ecology* 45: 252–258.

Ruehlmann, J. and M. Körschens. 2009. Calculating the effect of soil organic matter concentration on soil bulk denisty. *Soil Science Society of America Journal* 73: 876–885.

Schlesinger, W. H. and E. S. Bernhardt. 2013. *Biogeochemistry: An Analysis of Global Change*. Elsevier, Inc, Waltham, MA.

Schmidt, M. W. I. et al. 2011. Persistence of soil organic matter as an ecosystem property. *Nature* 478: 49–56.

Schothorst, C. 1977. Subsidence of low moor peat soils in the western Netherlands. *Geoderma* 17: 265–291.

Sinsabaugh, R. L. 2010. Phenol oxidase, peroxidase and organic matter dynamics of soil. *Soil Biology and Biochemistry* 42: 391–404.

Sinsabaugh, R. L., R. K. Antibus, and A. E. Linkins. 1991. An enzymic approach to the analysis of microbial activity during plant litter decomposition. *Agriculture, ecosystems & environment* 34: 43–54.

Smemo, K. A. and J. B. Yavitt. 2011. Anaerobic oxidation of methane: An underappreciated aspect of methane cycling in peatland ecosystems? *Biogeosciences* 8: 779–793.

Soil Science Society of America. 2008. *Glossary of Soil Science Terms*. Soil Science Society of America, Inc, Madison, WI.

Soil Survey Staff. 2004. *Soil Survey Laboratory Methods Manual, Soil Survey Investigations Report No. 42, Version 4.0*. USDA-Natural Resources Conservation Service, Washington, DC.

Soil Survey Staff. 2010. *Keys to Soil Taxonomy*, 11th edition. USDA-Natural Resources Conservation Service, Washington, DC.

Soil Survey Staff. 2011. *Soil Survey Laboratory Information Manual. Soil Survey Investigations Report No. 45, Version 2.0*. U.S. Department of Agriculture, Natural Resources Conservation Service.

Sposito, G. 2008. *The Chemistry of Soils*. Oxford University Press, Inc, New York, NY.

Stalheim, T., S. Ballance, B. E. Christensen, and P. Granum. 2009. Sphagnan—a pectin-like polymer isolated from Sphagnum moss can inhibit the growth of some typical food spoilage and food poisoning bacteria by lowering the pH. *Journal of applied microbiology* 106: 967–976.

Stevenson, F. J. 1994. *Humus Chemistry: Genesis, Composition, Reaction*. John Wiley & Sons, Inc, New York, NY.

Sutton, R. and G. Sposito. 2005. Molecular structure in soil humic substances: The new view. *Environmental Science & Technology* 29: 9009–9015.

Swift, R. S. 1996. Organic matter characterization. In D. L. Sparks (Ed.) *Methods of Soil Analysis: Part 3—Chemical Methods*. Soil Science Society of Americ and American Society of Agronomy, Madison, WI, pp. 1011–1069.

Szumigalski, A. R. and S. E. Bayley. 1996. Net above-ground primary production along a bog-rich fen gradient in central Alberta, Canada. *Wetlands* 16: 467–476.

Tarnocai, C., J. G. Canadell, E. A. G. Schuur, P. Kuhry, G. Mazhitova, and S. Zimov. 2009. Soil organic carbon pools in the northern circumpolar permafrost region. *Global biogeochemical cycles* 23: GB2023, doi: 10.1029/2008GB003327.

Tarnocai, C., I. M. Kettles, and B. Lacelle. 2005. *Peatlands of Canada*. Agriculture and Agri-Food Canada, Research Branch, Ottawa, ON, Canada.

Thormann, M. N., A. R. Szumigalaski, and S. E. Bayley. 1999. Aboveground peat and carbon accumulation potentials along a bog-fen-marsh wetland gradient in souther boreal Alberta, Canada. *Wetlands* 19: 305–317.

Tian, H. et al. 2010. Spatial and temporal patterns of CH_4 and N_2O fluxes in terrestrial ecosystems of North America during 1979–2008: Application of a global biogeochemistry model. *Biogeosciences* 7: 2673–2694.

Turetsky, M., W. Donahue, and B. Benscoter. 2011. Experimental drying intensifies burning and carbon losses in a northern peatland. *Nature Communications* 2: 514.

Turetsky, M., K. Wieder, L. Halsey, and D. Vitt. 2002. Current disturbance and the diminishing peatland carbon sink. *Geophysical Research Letters* 29: 21-1–21-4.

Verry, E. S., D. H. Boelter, J. Päivänen, D. S. Nichols, T. Malterer, and A. Gafni. 2011. Physical properties of organic soils. In R. K. Kolka, S. D. Sebestyen, E. S. Verry, and K. N. Brooks (Ed.) *Peatland Biogeochemsitry and Watershed Hydrology at the Marcell Experimental Forest.* Taylor & Francis Group, LLC, Boca Raton, Florida.

Vile, M. A., S. D. Bridgham, and R. K. Wieder. 2003a. Response of anaerobic carbon mineralization rates to sulfate amendments in a boreal peatland. *Ecological Applications* 13: 720–734.

Vile, M. A., S. D. Bridgham, R. K. Wieder, and M. Novak. 2003b. Atmospheric sulfur deposition alters pathways of gaseous carbon production in peatlands. *Global Biogeochemical Cycles* 17: 1058.

Vitt, D. H., L. A. Halsey, C. Campbell, S. E. Bayley, and M. N. Thormann. 2001. Spatial patterning of net primary production in wetlands of continental western Canada. *Ecoscience* 8: 499–505.

Watts, A. C. 2013. Organic soil combustion in cypress swamps: Moisture effects and landscape implications for carbon release. *Forest Ecology and Management* 294: 178–187.

Weishampel, P., R. Kolka, and J. Y. King. 2009. Carbon pools and productivity in a 1-km^2 heterogeneous forest and peatland mosaic in Minnesota, USA. *Forest Ecology and Management* 257: 747–754.

Weltzin, J. F., J. Pastor, C. Harth, S. D. Bridgham, K. Updegraff, and C. Chapin. 2000. Response of bog and fen plant communities to warming and water table manipulations. *Ecology* 81: 3464–3478.

Westerling, A. L., H. G. Hidalgo, D. R. Cayan, and T. W. Swetnam. 2006. Warming and earlier spring increase western US forest wildfire activity. *Science* 313: 940–943.

Ye, R., Q. Jin, B. Bohannan, J. K. Keller, and S. D. Bridgham. 2014. Homoacetogenesis: An underappreciated carbon pathway in peatlands. *Soil Biology & Biochemistry* 68: 385–391.

Yu, Z., J. Loisel, D. P. Brosseau, D. W. Beilman, and S. J. Hunt. 2010. Global peatland dynamics since the Last Glacial Maximum. *Geophysical Research Letters* 37: L13402.

Yu, Z. C. 2012. Northern peatland carbon stocks and dynamics: A review. *Biogeosciences* 9: 4071–4085.

Zinder, S. H. 1993. Physiological ecology of methanogens. In J. G. Ferry (Ed.) Methanogenesis. Chapman & Hall, United Kingdom, pp. 128–206.

7

Morphological Features of Hydric and Reduced Soils

Michael J. Vepraskas and Karen L. Vaughan

CONTENTS

Introduction

Hydric soils are created by oxidation–reduction (redox) chemical reactions that occur when a soil is anaerobic and chemically reduced. The redox reactions produce signs in the soil that they have occurred, and these signs are described in this chapter as *morphological features of reduction* and *hydric soil field indicators* (USDA NRCS 2010). A "reduced" soil is one in which redox reactions have caused reduced forms of O, N, Mn, Fe, or S to be present in the soil solution. "Reduced" is a general term that implies that some elements in addition to O_2 are present in their reduced form. Common reduced forms of elements or compounds that are found in hydric soils include H_2O, N_2, Mn^{2+}, Fe^{2+}, and H_2S, while their oxidized counterparts are O_2, NO_3^{2-}, MnO_4, FeOOH, and SO_4^{2-}, respectively. This chapter will focus on morphological features that form in soils that have been reduced periodically or seasonally. Hydric soil field indicators are the subject of Chapter 8.

Morphological features of seasonally reduced soils include specific color patterns, odors, color changes that occur on exposure to air, or a specific kind of organic material. These features can occur at any depth in the soil, and the abundance of a given feature is variable. The features are direct indicators that the soil was reduced at some point in its history, and therefore, they will be referred to as morphological features of reduction in this chapter. Morphological features of reduction have also been used to estimate which part of the soil is seasonally saturated with free water (Franzmeier et al. 1983; Cogger and Kennedy 1992).

On the other hand, most hydric soil field indicators are soil layers with precisely defined colors, thicknesses, and depths that contain morphological features of reduction in specific amounts. As their name suggests, the field indicators were developed solely to identify hydric soils on-site. Morphological features of reduction are components of hydric soil field indicators, but the two terms are not interchangeable. For example, when Fe hydroxides accumulate around root channels in sufficient quantities to be visible, they form the morphological feature of reduction called a Fe *pore lining*. If these pore linings occupy 3% of a sandy loam soil layer whose matrix has a Munsell color of 5YR 3/1, is 10 cm or more thick, and lies entirely within the upper 30 cm of the soil, then the layer qualifies as a hydric soil field indicator termed a *redox dark surface* (USDA NRCS 2010). A soil that contains morphological indicators of reduction may not be a hydric soil if the features occur too deeply; yet, they still indicate that the soil has experienced chemical reduction at some point in its history.

The purpose of this chapter is threefold: (i) to discuss the principal redox reactions and soil conditions needed to form morphological features of reduced soils and field indicators of hydric soils; (ii) to identify the principal types of morphological features of reduction and review their formation; and (iii) to discuss the ways these features can be interpreted. Additionally, the chapter discusses the relationship of hydric soil morphologies with soil texture, formation time, constructed wetlands, ditched areas, relict features, and problem situations.

Important Chemical Reactions

Principal Elements Involved

To understand how morphological features of reduction form, it is useful to simplify oxidation and reduction processes and consider them to be separate reactions even though

TABLE 7.1

Major Reducing Reactions Related to the Development of Hydric Soil Morphological Features

Reducing Reaction	Approximate Eh (pH 7)[a] (mV)	Morphological Feature Formed	
		Group Name	Examples
$O_2 + 4e^- + 4H^+ \rightarrow 2H_2O$[b]	600	Organic- C-based features	Oe, Oa, and some black A horizons
$MnO_2 + 2e^- + 4H^+ \rightarrow Mn^{2+} + 2H_2O$	300	Mn-based features	Mn concentrations and some depletions (black and gray mottles)
$2FeOOH + 4e^- + 6H^+ \rightarrow 2Fe^{2+} + 4H_2O$	100	Fe-based features	Fe concentrations and Fe depletions (red, yellow, and gray mottles)
$SO_4^{2-} + 8e^- + 10H^+ \rightarrow H_2S + 4H_2O$	– 200	S-based features	Odor of rotten eggs

[a] Data from McBride, M. B. 1994. *Environmental Chemistry of Soils.* Oxford University Press, New York.
[b] This reaction occurs under aerobic conditions, and it is not until the O_2 is depleted that organic-C accumulates.

the two types of reactions occur simultaneously in complex biochemical processes. As explained in Chapter 4, oxidation–reduction reactions in most soils begin when microorganisms oxidize organic compounds to release electrons and protons in the form of H^+ cations. The electrons and protons released by the oxidation of organic compounds react with electron acceptors to complete the microbial respiration process.

The principal reducing reactions that form the morphological features of reduction involve four elements (O, Mn, Fe, and S), which are used as electron acceptors in anaerobic microbial respiration. As shown in Table 7.1, there are also four basic groups of features that are associated with the principal reducing chemical reactions. These feature groups will be identified by the major element related to their formation: (a) organic-C-based features, (b) Mn-based features, (c) Fe-based features, and (d) S-based features. Selected examples of morphological features of reduction are given for illustration in Table 7.1. The reactions for Mn and Fe are reversible and produce different features as the reactions proceed in either direction.

There is a tendency among some wetland scientists to consider one of the four groups shown in Table 7.1 as inherently better or more reliable in identifying soils that have been reduced. This tendency must be avoided, because the preferred group is usually only the one with which they are most familiar. Each group of features is equivalent in showing that reducing reactions have occurred in the soil.

It is useful to place features in these groups because some soils are more likely to have one group of features than others. For example, some sands have virtually no Fe in them because the minerals found in the parent material simply did not contain Fe. The morphological features of reduction that will be found in such soils will consist of organic-C-based features, Mn-based features, and occasionally the S-based feature. It makes no sense to search these Fe-poor soils for Fe-based signs of redox reactions. Remember that the features in the four groups are equivalent in showing that reduction has occurred in the soil.

Relation of Features to Redox Potential

The formation of one of the groups of features shown in Table 7.1 requires that the redox potential falls to a certain Eh value. When an aerated soil becomes saturated, the reducing

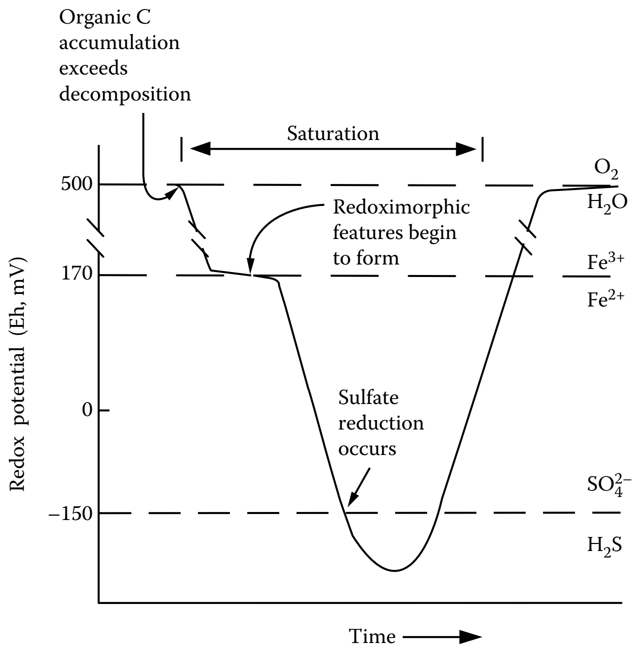


FIGURE 7.1
Hypothetical changes in redox potential (Eh) over the course of a single "wet" season where the water table rises, oxygen becomes depleted, and reduced soil conditions are induced. The water table then falls, causing the soil to reoxidize. The morphological features of reduction that would be expected to form at each change of Eh are also shown. For this scenario to occur, the soil must contain a respiring microbial community that is oxidizing organic-C materials.

reactions proceed in the order shown and progress from a higher Eh to lower Eh. An example of Eh fluctuation over a portion of the year is shown in Figure 7.1 to illustrate the relationship between seasonal Eh fluctuations and feature formation. When the soil is unsaturated and molecular O_2 is present in soil pores, the Eh is relatively high (>500 mV at soil pH 7), and none of the morphological features related to reduction shown in Table 7.1 is forming. Soils in this condition are described as oxidized or aerated. When the soils saturate to the surface, the movement of molecular O_2 from the atmosphere into the soil is retarded. Diffusion of O_2 into water is 10,000 times slower than through air (Chapter 3). Microbes that are still respiring by oxidizing organic compounds can reduce the remaining dissolved O_2 in the water. As shown in Figure 7.1, once all the dissolved O_2 has been depleted, the redox potential falls below 500 mV, and microorganisms must use other electron acceptors to survive. Morphological features formed by the buildup of organic material can begin to develop at this point because decomposition of organic tissues slows down under anaerobic conditions (Chapter 6).

The Eh continues to fall as long as saturation and anaerobic conditions are maintained and microorganisms continue to respire. When the Eh reaches approximately 170 mV (pH 7), the insoluble Fe^{3+} ions in some minerals will reduce to soluble Fe^{2+} and dissolve into the soil solution. The soluble Fe^{2+} may diffuse through the soil, concentrating in some areas or moving out of the soil horizon. As long as the Eh stays below 170 mV, the Fe^{2+} will remain reduced in most cases. The immediate change in the soil that occurs following Fe reduction is that the portion of the soil where reduction occurred will become grayer in color.

The gray color occurs because Fe^{2+} is colorless, and the actual color of the soil is determined by the color of the sand, silt, and clay particles in it.

Once all the Fe has been reduced, the Eh will continue to fall, and when it reaches –150 mV, SO_4^{2-} anions may be reduced to H_2S gas. This usually requires a relatively long period of saturation and anaerobic respiration. The gas is produced only while the Eh is below –150 mV.

When the soil drains, O_2 reenters the soil and the Eh increases. The production of H_2S ceases, and reduced Fe is oxidized to Fe oxide or hydroxide minerals, which produce the red, yellow, or brown colors seen in many subsoil horizons. Above an Eh of 500 mV, aerobic organisms respire and oxidize undecomposed tissues to CO_2 and water.

Basic Kinds of Features

The widespread features that have been found in reduced soils will be described for each of the principal groups of features shown in Table 7.1. Additional information on features related to organic matter are included in Chapters 6 and 8, while those related to H_2S gas formation are discussed in Chapters 4 and 13.

Organic-C-Based Features

All organic-C-based features consist of one of the three kinds of materials that were defined in Chapter 1: organic soil material, mucky mineral soil material, or mineral soil material with a black color. These materials form either distinct horizons (O or A horizons) or occur as aggregates of organic-rich material. O horizon thickness and the state of decomposition must be considered when identifying hydric soils. O horizons having a thickness of >20 cm and a black or very dark gray color, regardless of the state of decomposition, are in most cases found only in soils that were periodically reduced (USDA NRCS 2010). Thinner layers can also be found in seasonally reduced soils, but their thickness and state of decomposition requirements vary for different textural groups and land resource regions, as described in Chapter 8. A horizons consist of mineral (occasionally mucky mineral) soil material, and those that formed in reduced soils have moist Munsell colors with a value of 3 or less and a chroma of 3 or less. The dark color is a direct result of relatively high amounts of organic matter that accumulated under reduced conditions. However, not all soils having dark colors necessarily experience periods of saturation and reduction. This is particularly true of the Mollisols found in the midwestern United States (Bell and Richardson 1997).

Another group of organic-C-based morphological features related to reduction are aggregates of organic materials called *organic bodies* that form around roots. They may be found in mineral horizons or within or just below O horizons located near the surface. These features consist of organic material or mucky mineral materials as described in Chapter 8.

Iron and Manganese-Based Morphological Features Related to Reduction

Redoximorphic features are formed by the reduction, movement, and oxidation of Fe and Mn compounds. These features form the gray, red, yellow, brown, or black-mottled color patterns that are normally associated with saturated and reduced soils. Redoximorphic features are the most widely observable morphological features formed by reduction. There are three basic kinds of redoximorphic features: *redox concentrations*, *redox depletions*, and the *reduced matrix*.

Redox reactions affect Fe and Mn similarly, and the two elements frequently occur together, as noted in Chapter 4. Iron is usually in greater abundance than Mn, but small quantities of Mn can cause some redox concentrations to appear black (Gallaher et al. 1974; Rhoton et al. 1993). Black-colored redox concentrations may be confused with the decomposed organic tissue. However, Mn can be detected by spraying the redox concentration with a weak (3% concentration) solution of hydrogen peroxide (H_2O_2). A rapid bubbling of the H_2O_2 solution confirms that Mn is present. The reaction is

$$MnO_4 + H_2O_2 \rightarrow Mn(OH)_2 + 2O_2 \tag{7.1}$$

where Mn is chemically reduced by the peroxide (Jackson 1969). While a 3% solution of H_2O_2 will react with soil organic matter, the reaction is slower than for Mn, although it can be sped up by heating. Manganese can be abundant in certain soils, such as those having pHs >7, or in some clays having Munsell hues of 5YR or redder (e.g., Moreland series reported in Hudnall et al. 1990). When Mn is abundant, it can prevent the reduction of Fe and formation of gray soil colors because it is reduced before Fe (McBride 1994). Such Mn-rich soils are probably of small extent, but can be important in certain regions. The remainder of this chapter will focus on Fe, but Mn should be assumed to be included as well.

Redox Concentrations

Redox concentrations are features formed when Fe oxides or hydroxides have accumulated at a point or around a large pore such as a root channel. They have been defined as "bodies of apparent accumulation of Fe–Mn oxides and hydroxides" (Vepraskas 1996, p. 7). This means that they appear to have formed by Fe or Mn moving into an area, oxidizing, and precipitating. Redox concentrations contain more Fe^{3+} oxides and hydroxides than were found in the soil matrix originally. Three kinds of redox concentrations have been defined: Fe masses, Fe pore linings, and Fe nodules and concretions. These differ in their hardness and also in where they occur in the soil.

Iron masses (Figure 7.2) are simply soft accumulations of Fe^{3+} oxides and hydroxides that occur in the soil matrix, away from cracks or root channels. They can be of any shape. The masses are soft and easily crushed with the fingers because the concentration of Fe is not great enough to cement the soil particles into a solid mass. Sizes of Fe masses range from 1 mm to more than 15 cm in diameter. Because they are found in the matrix, the size of the Fe masses is usually determined by the size of the peds or structural aggregates in the soil that fix the maximum size for the features.

The color of the Fe masses is variable and can be any shade of red, orange, yellow, or brown. The color varies with the type and concentration of Fe mineral present. The most common Fe minerals found in Fe masses are goethite, ferrihydrite, and lepidocrocite (Schwertmann and Taylor 1989). These minerals impart hues of 10YR, 7.5YR, and 5YR, respectively. Common value/chroma combinations include 5/6 and 5/8, but other combinations can be found.

Pore linings (Figures 7.3 and 7.4) are accumulations of Fe oxides and hydroxides that lie along ped surfaces or root channels. These features occur in the soil and not directly on the root. They are similar to oxidized rhizospheres, but while oxidized rhizospheres are

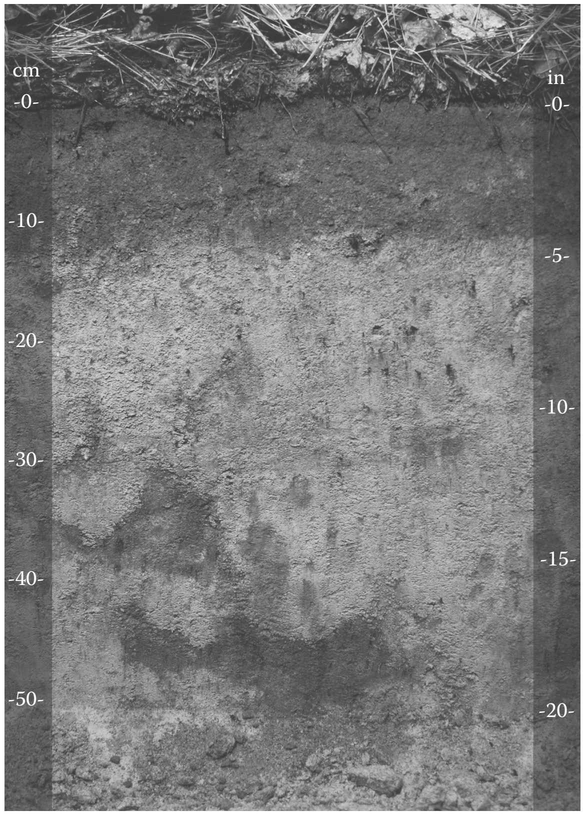


FIGURE 7.2
(**See color insert.**) Example of iron masses (reddish orange colors) between 30 and 50 cm. (Photo provided by John Kelley.)

thought to form on the root tissue when the root is alive (Mendelssohn et al. 1995), pore linings do not need a live root to form. The distinction between pore linings and oxidized rhizospheres is not important for identifying hydric soils. However, if one needs to identify indicators of wetland hydrology, which currently requires the soil to be periodically saturated during the growing season when plants are growing (Environmental Laboratory 1987), then, only oxidized rhizospheres occurring along living roots can be used because pore linings could develop outside the growing season when soils are reduced and become oxidized as the water table falls (Megonigal et al. 1996).

Pore linings differ from iron masses only in where they occur in the soil: masses occur in the matrix, while the pore linings must be along root channels or cracks. The colors of the two features are similar. Pore linings are generally soft, but in extreme cases, the Fe content has reached a level that cements the soil particles together around a root channel. The cemented feature has been called a *pipestem* because it is usually cylindrical and has a small channel running down its axis resembling the shaft of a smoker's pipe (Bidwell et al. 1968).

Nodules and concretions (Figure 7.5) are hard, generally spherical-shaped bodies made of soil particles cemented by Fe oxides or hydroxides. They range in size from less than 1 mm to more than 15 cm in diameter. When broken in half and examined, the concretions are seen to consist of concentric layers such as an onion, while no layers are seen in nodules.

FIGURE 7.3
(**See color insert.**) Example of an iron pore lining (10 mm wide) along a root channel. (Photo provided by John Kelley.)

Most people seem to use the two terms interchangeably, and there is no special significance attached to the layered structure other than it shows that the concretion formed in episodes over time.

The nodules and concretions are difficult to destroy because of their hardness. When they are found in soils, it is never clear whether these features formed in place or were brought into the soil by flooding or by deposition of material eroded from upslope. For this reason, nodules and concretions cannot be considered as reliable indicators of the processes that still occur seasonally in the soil.

Redox Depletions

Redox depletions are zones formed by loss of Fe and other components. They have been defined as "bodies of low chroma (<2), having values of 4 or more where Fe–Mn oxides alone have been stripped out or where both Fe–Mn oxides and clay have been stripped out" (Vepraskas 1996, p. 10). This definition is used to meet the soil classification requirements for aquic conditions as set forth in *Soil Taxonomy* (Soil Survey Staff 2010). Redox depletions in principle could form with chromas >2 as long as they developed in a soil horizon whose matrix lost Fe by reduction processes.

Two different kinds of redox concentrations have been defined, *Fe depletions* (Figures 7.6 and 7.7) and *clay depletions*, and these differ only in whether their texture is similar to that of the matrix or not. Iron depletions simply form by a loss of Fe (and/or Mn) from a portion of the soil. They have been defined as "low chroma bodies (chromas <2) with clay content

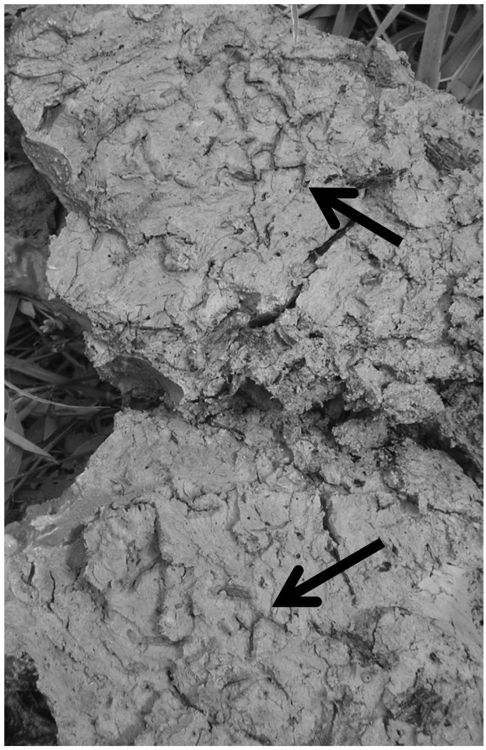


FIGURE 7.4
(**See color insert.**) Examples of iron pore linings (5–10 mm wide) in a horizon containing a Depleted Matrix hydric soil field indicator. (Photo provided by M. Vepraskas.)

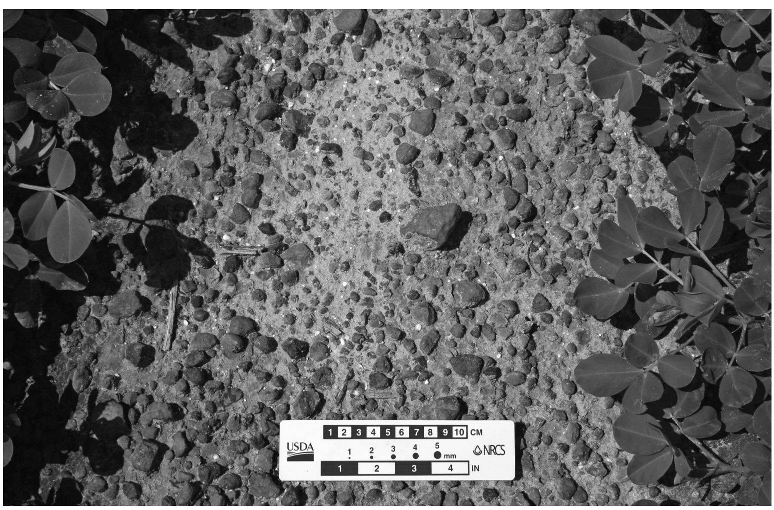


FIGURE 7.5
(**See color insert.**) Iron nodules on the soil surface that were exposed by erosion. (Photo provided by John Kelley.)

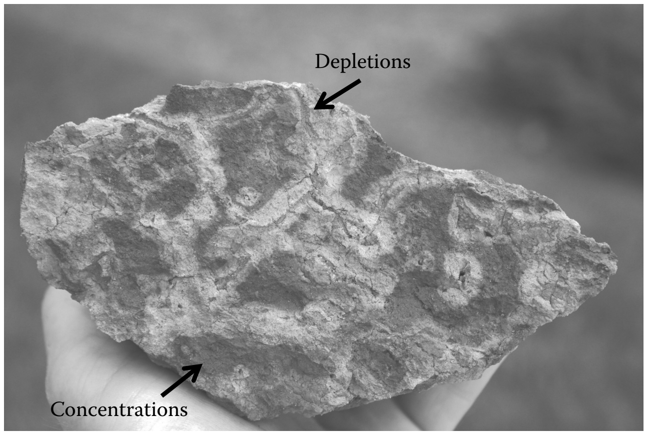


FIGURE 7.6
(**See color insert.**) Gray iron depletions (10 mm wide) along root channels. Reddish orange iron masses in the matrix. (Photo provided by John Kelley.)

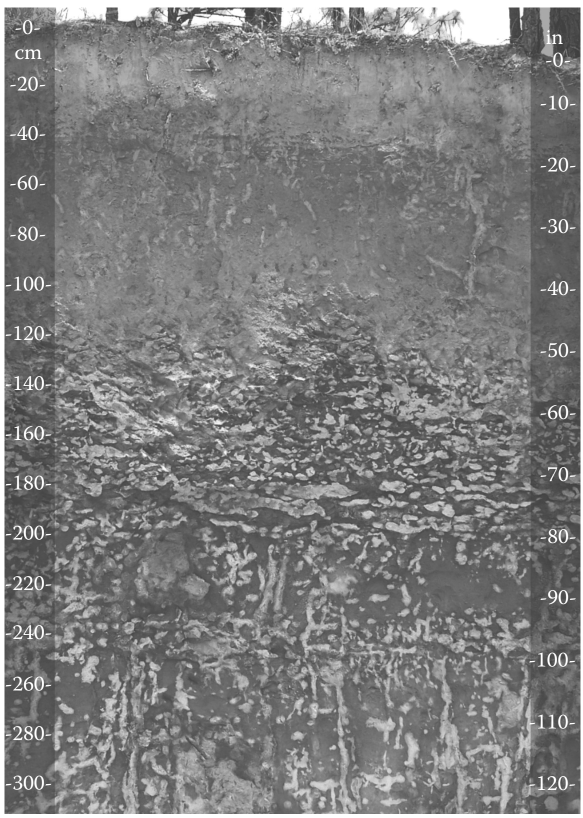


FIGURE 7.7
(**See color insert.**) Iron depletions along root channels primarily below a depth of 120 cm in a Fragic Kandiudult soil in NC. (Photo provided by John Kelley.)

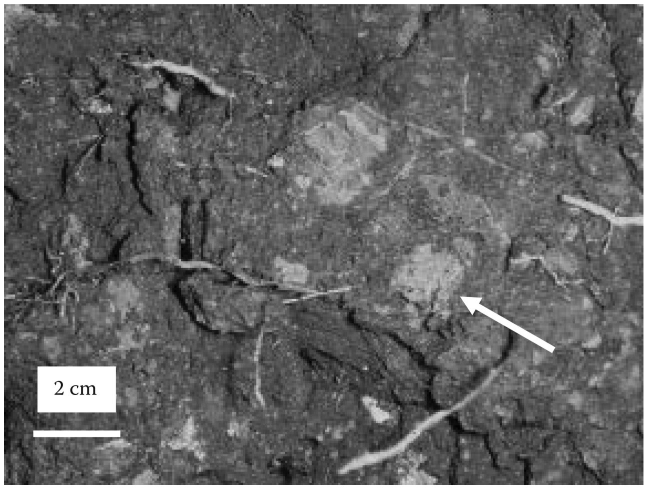


FIGURE 7.8
(**See color insert.**) Gray iron depletions (arrow) in matrix of a silt loam A horizon. (Photo provided by M. Vepraskas.)

similar to that of the adjacent matrix" (Vepraskas 1996, p. 10). Similar features have also been called gray mottles, gley mottles, albans, and neoalbans (Veneman et al. 1976). Iron depletions frequently occur along root channels and ped surfaces in B horizons. They can occur in the matrix, particularly in A horizons as shown in Figure 7.8. In some cases, the entire matrix is an Fe depletion, such as in E and B horizons of soils that are reduced for long periods.

Clay depletions form by a loss of both Fe and clay. These features have both a low chroma and a coarser texture than found in the adjacent soil matrix. Clay depletions have also been described as silt coatings, skeletans, and neoskeletans (Brewer 1964; Vepraskas and Wilding 1983). They almost always occur along ped surfaces or large root channels. Similar features that occur in the soil matrix were probably formed by silt or sand falling down into and filling a channel. These are technically not clay depletions because their formation is not related to oxidation–reduction chemical reactions. Clay depletions have not been reported within the upper 30 cm of hydric soils and are less important than Fe depletions for hydric soil identification.

Reduced Matrix

The reduced matrix has been defined as a soil matrix that has a "low chroma color *in situ* because of the presence of Fe^{2+}, but whose color changes in hue or chroma when exposed to air as the Fe^{2+} is oxidized to Fe^{3+}" (Vepraskas 1996, p. 10). The time needed for the color change to occur was set at 30 min for practical reasons, because waiting longer would interfere with the completion of field work. However, there are little data on how long such color changes require.

In principle, the reduced matrix could also be detected by spraying a field moist sample of soil with a dye such as α,α'-dipyridyl that reacts with Fe^{2+}. If a positive reaction with the dye occurs, then, it might be assumed that a reduced matrix is also present (Griffin 2008). The developers of *Soil Taxonomy* (Soil Survey Staff 2010) incorporated this technique as a substitute for redox concentrations if no such features are present. It is also a test to show that the soil is reduced.

Soils that have a reduced matrix during the wet season may develop other morphological features during other periods of the year when the soils drain and oxygen enters some pores. At the time that the reduced matrix occurs in a soil, it is likely that some or all the redox concentrations (especially pore linings and Fe masses) that were present in the soil prior to the time it became reduced will have dissolved. The dissolved redox concentrations may reform after the water table falls and the Fe^{2+} cations are oxidized. Some hydric soil field indicators, such as the redox dark surface (USDA NRCS 2010), require, in addition to a black matrix color, that a certain percentage of redox concentrations be present before the indicator is met. If the soil is examined at a time when only the reduced matrix is present, the soil may not meet a hydric soil field indicator even though it is actually saturated and reduced at the time of observation. To overcome this problem, it is recommended that a positive test for Fe^{2+} be used as a substitute for redox concentrations in soil descriptions, particularly when hydric soils are being examined.

Gley Soil Colors

Gley colors as defined in Chapter 1 can be one of the two types of redoximorphic features, either a reduced matrix or depleted matrix, depending on whether the color changes when exposed to air. Gley colors consist of Munsell hues found on the "Gley Pages" of a Munsell Color Chart. In addition, soils with a gleyed matrix can possess the following combination of hue, value, and chroma: 10Y, 5GY, 10GY, 10G, 5BG, 10BG, 5B, 10B, or 5PB with a value of 4 or more and chroma of 1; 5G with a value of 4 or more and chroma of 1 or 2; or N with a value of 4 or more. If the color of the soil does not change upon air drying, the material with the gley hue is probably a redox depletion (Fe depletion). This occurs when the Fe is reduced, translocated, and therefore removed or transformed so that it is no longer present in the soil matrix. If a color change (e.g., reddening of the matrix) occurs upon drying, the material is probably a reduced matrix even though the color change did not occur within 30 min of the sample being removed from the soil. A reduced matrix indicates that the Fe is reduced but not translocated from the soil matrix so that, upon exposure to oxygen, the iron is oxidized to a colored form.

The gley hues can, in some cases, be unique minerals that contain a reduced form of Fe that is combined with an anion of phosphate, sulfate, carbonate, or other compounds (Schwertmann and Taylor 1989). These hues are a morphological indicator of reduction when their values are 4 or more. When values are less than 4, the colors are probably too dark to accurately separate gley hues from other kinds.

Formation of Redoximorphic Features

Formation of the organic features related to reduction was described in detail in Chapter 6 and the production of H_2S was described in Chapters 4 and 5, and these processes will not be repeated here. This section will focus only on the formation of redoximorphic features. Redoximorphic features form after one or more of the following three processes have occurred: (1) Fe^{3+} cations in oxides or hydroxides have been reduced, (2) the solubilized Fe^{2+} has moved to another portion of the soil, and (3) the Fe^{2+} has been oxidized to form an Fe mass, pore lining, or nodule. A reduced matrix requires that only the first step occurs. Iron depletions require that the first two steps occur, while the formation of redox concentrations requires that all three steps occur.

Redox Depletions

The process of redox depletion formation is shown in Figure 7.9 for an Fe depletion. To begin the process, assume that the soil matrix had a uniform brown color throughout before an Fe depletion formed. The color came from Fe^{3+} oxide or hydroxide minerals that coated soil particles, such as sand, silt, and clay grains. Each of these grains had a coating of an Fe^{3+} compound that effectively painted the particle surface brown.

The soil shown in Figure 7.9a has a channel containing a dead root that is being decomposed by bacteria. If the channel is filled with air, the oxidation of the organic matter releases electrons that are used to reduce O_2 to water. As long as the channel contains air, the only reduction that takes place is that of O_2 to water, and there is no change in the color of the soil around the root.

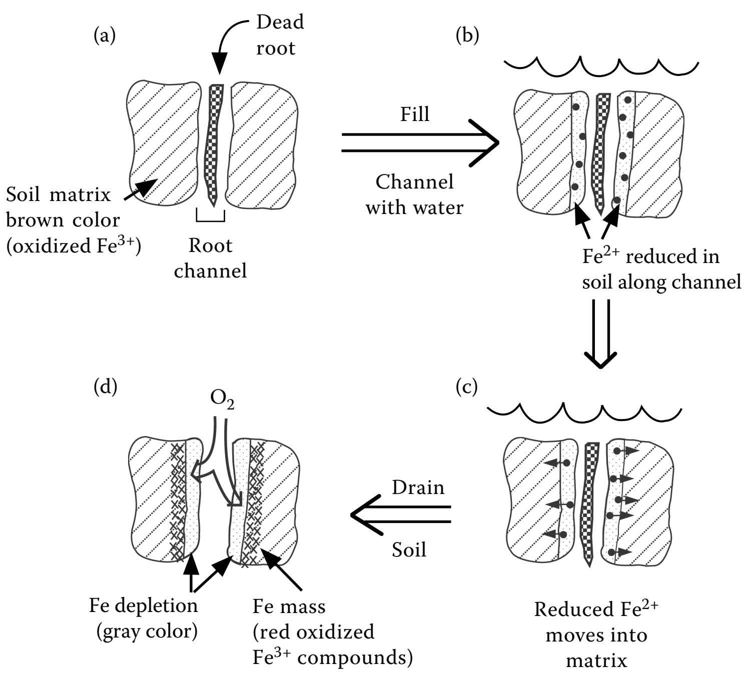


FIGURE 7.9
Formation of redox depletions (Fe depletions) around a root channel. Initially (a) the soil is uniformly brown throughout its matrix. The channel contains a dead root, which is being decomposed by bacteria. Upon flooding (b) the channel is filled with water, and all O_2 dissolved in the water is reduced. When the water is anaerobic, the bacteria reduce Fe oxides and hydroxides in the soil surrounding the channel. The Fe^{2+} dissolves and moves away from the channel (c) leaving the soil particles around the channel stripped of the Fe coatings. If the stripped grains are largely composed of quartz and/or uncoated clay minerals, they will be gray in color (d). The Fe^{2+} may subsequently oxidize to form a reddish-brown Fe mass. (Adapted from Vepraskas, M. J. 1996. *Redoximorphic Features for Identifying Aquic Conditions.* Technical Bulletin 301. North Carolina Agricultural Experiment Station, Raleigh, NC.)

If the channel fills with water, however, the supply of O_2 from the atmosphere is cut off (Figure 7.9b). Microbial respiration still occurs, and the organic tissue continues to be oxidized. The O_2 dissolved in the soil water is quickly depleted because it is the primary electron acceptor used in the respiration process. After the water becomes anaerobic, the bacteria must use another element to accept electrons to sustain their respiration process. As noted in Chapter 4, the alternative electron acceptor elements used for reduction occur in a sequence, but for this chapter, we will focus on Fe.

When the electrons are transferred to Fe^{3+} atoms in minerals, the reduction of Fe causes two changes in the minerals: they lose their color and the Fe^{2+} dissolves. The dissolved Fe^{2+} moves off the particle surfaces and may diffuse through the soil matrix or may be carried to other parts of the soil in moving water (Figure 7.9c). As Fe^{2+} leaves the particle surfaces, the color of the soil around the channel changes and the soil gradually becomes gray in color. In Munsell terminology, the soil chroma decreases and value increases until all Fe has been removed from the surfaces of particles lying near the channel. When the soil drains and O_2 enters the soil, the newly formed Fe depletion will retain its gray color because it is the color of the uncoated mineral soil grain. Oxygen penetrating into the matrix may oxidize the Fe^{2+} and cause the formation of pore linings or Fe masses (Figure 7.9d).

Stripping the soil particles of Fe^{3+} mineral coatings changes the soil's color to the natural color of the soil minerals. Normally, this is a gray color when the particles consist of quartz or a clay mineral such as kaolinite. This gray color is relatively permanent and is not affected appreciably by additional periods of saturation and reduction. The only way the color of the gray soil particles could become brown again is if Fe^{2+} were moved onto the particle surfaces and reoxidized to recoat the particle surfaces with more of the "paint" composed of Fe^{3+} minerals.

This is the basic process that forms redox depletions. It is easiest to see in the field when the organic matter occurs as a root that is not near other roots as in Figure 7.7. In such cases, the gray depletions form cylindrical features around root channels. However, if roots are closely spaced along a crack or ped surface, then, the depletions will have a planar shape as they coat the surface of the crack, or they may occupy entire layers.

In A horizons (Figure 7.8), the organic matter that starts depletion formation can be a piece of leaf tissue or a fragment of some other part of the plant. This is an isolated source of the organic tissue. The depletions that form around this tissue tend to be spherical. The process that forms the depletion is the same as that shown in Figure 7.9.

Redox Concentrations

Redox concentrations can occur both in the matrix as Fe masses or Fe nodules, and around macropores such as root channels in the case of Fe pore linings. Redox concentrations form when Fe^{2+} in the solution moves through the soil toward points of oxidation and precipitates. Points of oxidation can occur in reduced soil where (1) O_2 enters the soil after a soil drains, (2) when entrapped air is present, or (3) when roots release O_2 to the soil matrix when Fe^{2+} is present. Figure 7.9d illustrates a case where Fe^{2+} diffused into a soil matrix and oxidized where the Fe^{2+} encountered entrapped oxygen, which may occur as an air bubble in the saturated soil. Figure 7.10 illustrates two cases, the first (Figure 7.10a) being where points of oxidation occur around roots that have O_2 transported to them. This process is one way in which pore linings form. The second example (Figure 7.10b) illustrates a case in which O_2 penetrates along macropores such as cracks or root channels and forms both Fe pore linings and Fe masses as Fe^{2+} diffuses to points where O_2 occurs. The formation of Fe pore linings and masses illustrated above has been modeled in laboratory experiments

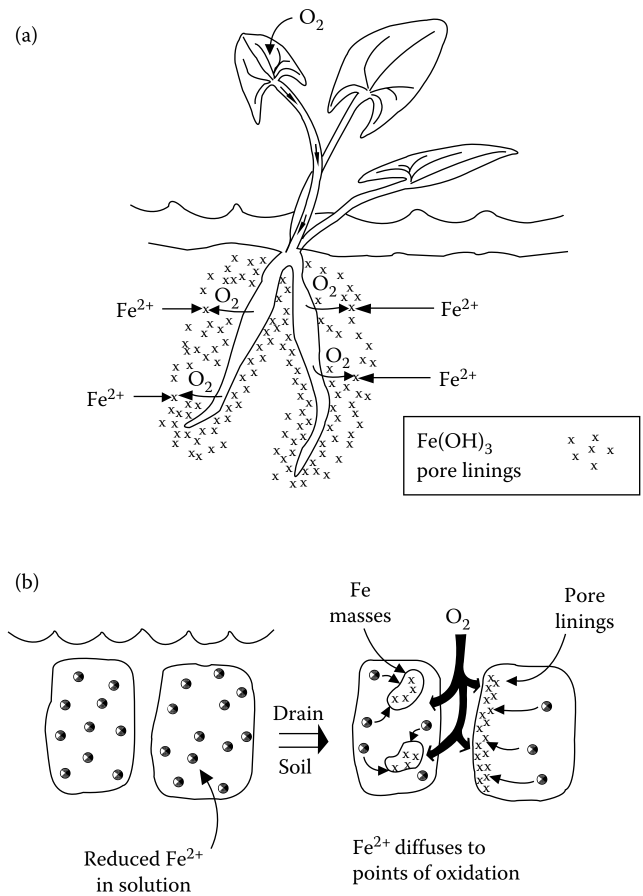


FIGURE 7.10
Formation of redox concentrations by two different processes. In (a) the Fe^{2+} is oxidized around roots that are bringing O_2 into the flooded and reduced soil. Pore linings are the type of redox concentration formed. In (b) the Fe^{2+} is oxidized in the matrix after the soil has drained, and O_2 has been able to penetrate into reduced portions of the matrix. Iron masses are formed in this case. It is also possible for Fe masses to form where air has been trapped inside peds. (Adapted from Vepraskas, M. J. 1996. *Redoximorphic Features for Identifying Aquic Conditions.* Technical Bulletin 301. North Carolina Agricutural Experiment Station, Raleigh, NC.)

that simulated field conditions (Vepraskas and Bouma 1976). To this author's knowledge, no experiments have simulated the formation of Fe nodules and concretions. Apparently, these features form slowly over time by repeated episodes of Fe oxidation at the same points in the soil, such as at the interiors of peds where oxygen is entrapped when the soil saturates.

Reduced Matrix

The reduced matrix occurs in soils by a process similar to that shown in Figure 7.9. Once reduced, the Fe^{2+} may remain in place or it may move to portions of the soil and concentrate. This author has seen reduced matrices along cracks in clay soils (Vertisols) where Fe^{2+} produced along a crack was not able to diffuse away from the crack. The time required to form a reduced matrix has not been determined.

FIGURE 7.11
(**See color insert.**) Gray iron depletions (arrows) in a loamy sand E horizon. Example of a Stripped Matrix hydric soil field indicator. Note dime for scale. (Photo provided by Wade Hurt.)

Effects of Texture on Redoximorphic Feature Appearance

Despite forming by similar processes, Fe depletions in sandy soils usually appear different from similar features formed in loamy or clayey soils. This can be seen by comparing the soils in Figures 7.7 and 7.11. Figure 7.7 shows a redox depletion around a root channel that formed by the process described in Figure 7.9 and Figure 7.11 shows a redox depletion in a sand that was formed by the same process but appears as a roughly circular gray area that is sometimes described as a "splotchy pattern." The Fe depletions that appear in both Figures 7.7 and 7.11 are formed by the process shown in Figure 7.9. They look different because sands do not have large, stable root channels or cracks that remain open for long periods to allow large features to form around the same points in the soil. Root channels in sand remain open while they contain a root, but collapse shortly after the root dies and decomposes. As a result, redox depletions develop around a single root, and their shape is determined by the arrangements of stripped sand grains (i.e., those free of Fe coatings and organic matter) that fell into the collapsing channel after the root decomposed. Furthermore, the low amounts of Fe oxides in sands cause the contrast between the matrix and the depletions to be less than in more Fe-rich loamy and clayey soils.

Time Needed to Form Redoximorphic Features

To become reduced, a soil must (1) be saturated with water to exclude O_2 from the atmosphere, (2) contain actively respiring microorganisms, and (3) be depleted of dissolved O_2. If any of these conditions are not met, the reduction of Fe will not occur. To achieve these conditions, an adequate supply of decomposable organic-C must be available, the soil water should be stagnant, and the soil temperatures must allow for microbial activity. If organic-C levels are too low, there may not be sufficient microbial respiration to deplete the soil water of oxygen even when the soil is saturated. Moving water tends to carry oxygen into the soil and retards the onset of Fe reduction (Cogger and Kennedy 1992; Gilman 1994). Furthermore, it is generally believed that at temperatures <5°C, microbial respiration will be too slow to deplete the soil water of oxygen (Megonigal et al. 1996). This 5°C threshold, also known as a biologic zero, is a general one that works best for plants whose roots have their maximum elongation rate at a soil temperature of 20–30°C (Russell 1977). The 5°C threshold is less applicable to organisms adapted to life in colder soils; therefore, a new concept of a biologic zero was introduced by the National Technical Committee for Hydric Soils and is tied to higher plants with deeper rooting depths (Rabenhorst 2005).

The time required for Fe reduction to occur after initiation of saturation or inundation depends on soil conditions. Meek et al. (1968) detected Fe^{2+} in solution after 1 day of ponding in field plots (1.2 by 1.2 m) to which chopped alfalfa had been added. The amount of Fe^{2+} in solution reached its peak of 5–30 mg/L at approximately 4–5 days following the initial ponding. Ponnamperuma (1972) reported that in acid soils "high in organic matter," the peak in Fe^{2+} occurs with 1–3 weeks of ponding. Other field studies have shown that Fe reduction may be delayed by up to 4 weeks following saturation, and may not occur at all, depending on soil conditions (Hayes and Vepraskas 2000).

The effect that organic matter content and temperature have on the time it takes for Fe reduction to occur is shown in Table 7.2 (Cogger and Kennedy 1992). These data show that there is a lag between the onset of saturation and the onset of Fe reduction, and that the length of the lag period depends on both soil temperature and organic matter percentage. These two factors directly influence the rate of microbial activity. The data in Table 7.2 illustrate why two soils that are saturated for the same length of time could develop widely different amounts of low chroma or gray color as a result of Fe reduction. For example, assume that two soils (represented by cores X and Y in Table 7.2) were saturated for 100 days each year. Further, assume that soil horizon X became saturated when its soil temperature was 23°C, while soil horizon Y became saturated when its soil temperature was 9°C. Iron reduction would be expected to last for 94 days in soil X, but for only 3 days in soil Y before the water tables fell in each soil. If the amount of gray color produced in each soil is directly proportional to the length of time they are reduced, then, we would expect the amount of gray color seen in horizon X to be more than 30 times that seen in horizon Y, despite both horizons being saturated for identical lengths of time.

The soluble Fe^{2+} can move through the soil with moving water or by diffusion. Vepraskas and Guertal (1992) modeled the formation of Fe depletions and found that diffusion is responsible for most of the Fe loss. This is because in many cases, water in wetland soils tends to be stagnant.

The oxidation of Fe^{2+} can occur quickly. In laboratory experiments, Ahmad and Nye (1990) showed that after 8 h, approximately 78% of the Fe^{2+} in both the solution and suspension

TABLE 7.2

Period Required for Saturated Soil Cores of Different Organic-C Percentages to Develop Fe-Reducing Conditions under Three Different Soil Temperatures

		Soil Temperature		
		23°C	9°C	4°C
Soil Core	Organic-C (%)	(days)		
X	7.5	6(1–20)[a]	37(22–95)	74(43–120)
Y	2.5	30(10–43)	97(40–140)	160(151–164)
Z	0.8	53(37–72)	97(80–147)	160(47–>180)

Source: Adapted from Cogger, C. G. and P. E. Kennedy. 1992. *Soil Sci.* 153(6): 421–433.

[a] Mean (range).

had oxidized at 20°C. Additional experiments with suspensions kept at a pH of 5.75 showed that approximately 60% of the Fe^{2+} had oxidized within 3 h. These results agree with field observations. For example, the reduced matrix is detected by a visible color change that is expected to occur within 30 min of exposure to air (Soil Survey Staff 2010). Movement of sufficient Fe^{2+} through the soil and its oxidation to form visible features has been found to occur around the roots of rice seedlings growing in a flooded field in 7 days (Chen et al. 1980).

The relationship between saturation time and the formation of redoximorphic features appears to vary widely depending on the geographic region, landscape position, soil type, and climate. Research has been done to relate the presence of redoximorphic features with the percent of time that a horizon is saturated (Daniels et al. 1971; Zobeck and Ritchie 1984a, b; Megonigal et al. 1993; Genthner et al. 1998; Szogi and Hudnall 1998; West et al. 1998; Jacobs et al. 2002; Fiedler and Sommer 2004; Morgan and Stolt 2006). For example, West et al. (1998) studied soils on the Georgia Coastal Plain and confirmed the presence of redox concentrations in soil horizons that were saturated for 20% of the time, redox depletions (2 chroma or less) in horizons that were saturated for approximately 40% of the time, and a depleted matrix in horizons that were saturated for about 50% of the time. In the coastal plain of North Carolina, Daniels et al. (1971) documented that soil horizons with gleyed matrices were saturated between 25% and 50% of the time, and soils with 3 chroma matrix colors (value of 6 or 7) were saturated for 25% of the time. Research by Franzmeier et al. (1983) supports the findings that soils with 3 chroma colors may be wetter than suggested in *Soil Taxonomy* (2010).

Constructed Wetlands

Rates of redoximorphic feature formation were studied under field conditions across the edge of a created deep marsh near Chicago, Illinois, by Vepraskas et al. (1999), and results are shown in Table 7.3. The hydrology of the site was controlled by pumping that brought water to the marsh. Soil horizons were described at three locations: in the marsh, at the edge of the marsh, and in a transition zone bordering the upland. The amount of redox depletions increased over the 5-year period (Table 7.3), but rapid changes occurred within the first 3 years. The transition zone developed more than 70% redox depletions within 3 years because the original soil matrix had a chroma of 3, and developing the redox depletions required losing enough Fe to produce a chroma of 2. Redox depletions decreased slightly by the 5th year because the water table had dropped in this transition zone after two relatively dry years. The data in Table 7.3 show that detectable changes in redox depletions occur quickly following changes in the soil hydrology.

A companion study to that of the deep marsh was conducted along a created floodplain adjacent to a constructed channel that had dams designed to control water entry and exit (Vepraskas et al. 2006). This allowed the number and duration of floods to be controlled. The topsoil applied to plots on the created floodplain was mixed and applied after the floodplain contours had been graded. The purpose of the study was to determine whether redox depletions could form in A horizons that were inundated by short-term floods. The soils were Mollic Endoaquents containing 2% organic-C in the A horizon.

Results of the study are shown in Table 7.4. The first induced flood lasted for 8 days and produced redox depletions like that shown in Figure 7.8 that occupied approximately 2% of the horizon's volume. Over the next 3 years, more depletions were formed as the number of floods increased. This study showed that such depletions can form even after a single inundation. On the other hand, after 3 years, the flooding stopped and the depletions began to disappear.

TABLE 7.3

Changes in the Quantity of Redox Depletions (Features Having Munsell Values of 4 or More and Chromas of 2 or Less) in Soils along and in a Created Deep Marsh

Soil Depth (cm)	Saturation (% of Year)	Organic-C (% by Weight)	Redox Depletions		
			Original Soil	After 3 Years	After 5 Years
			(% by Volume)		
Marsh					
18–25	100	1.1	50	90	88
25–58	100	0.5	50	80	89
Edge of Marsh					
10–30	100	1.3	0	40	75
30–53	100	0.8	50	50	70
Transition to Upland					
13–23	30[a]	1.5	0	85	70
23–41	30	1.0	0	75	60

Source: Adapted from Vepraskas, M. J. et al. 1999. *Wetlands* 19(1): 78–89.
Note: The hydrology was controlled by pumping.
[a] Estimated from bimonthly water table data.

Ditching Effects

Hayes and Vepraskas (2000) evaluated changes in soil morphology in a Coastal Plain landscape (interstream divide) at four different distances from a ditch. He found that after 30 years, the Bt horizons of soils within 30 m of the ditch had significantly (0.10 level) greater amounts of redox concentrations than soils farther from the ditch (Table 7.5). The near doubling of redox concentrations was related to the soils nearest the ditch being reduced for significantly shorter periods of time. Reduced Fe in groundwater flowing toward the ditch precipitated in the soils near the ditch. The duration of saturation was not affected by the ditch, because while the ditch removed groundwater from the soils within 30 m,

TABLE 7.4

Formation of Redox Depletions in the A Horizon of Soils along a Created Floodplain as a Function of Flood Frequency and Duration

	Year of Study		
	1992	1993	1994
Number of floods	2	5	2
Flood durations (days)	7–11	4–44	13–14
Redox depletion characteristics			
Abundance (%)	2	7	27
Color (moist)	2.5Y 4/1	5Y 4/1	2.5Y 4/1
Size (mm)	2–10	2–35	2–20

Source: Adapted from Vepraskas, M. J., J. L. Richardson, and J. P. Tandarich. 2006. *Wetlands* 26: 486–496.
Note: The soil was classified as a Typic Endoaquoll and A horizon had a pH of 7 with 2.3% organic-C.

TABLE 7.5

Effect of a Drainage Ditch on Durations of Saturation, Fe Reduction, and Redox Concentrations for a Typic Paraquat in the Coastal Plain Region of North Carolina

Distance from Ditch (m)	Duration of Saturation (% of Year)	Duration of Fe Reduction (% of Year)	Redox Concentrations (% by Volume)
7	41a[a]	13a	39a
30	44a	22a	38a
60	44a	39b	16b
80	45a	34b	20b

Note: Data for saturation and reduction were determined at 60 cm for 1 year, while the abundance of redox concentrations was determined for the depths between 40 and 60 cm by Hayes and Vepraskas (2000).

[a] Numbers within the same column that are followed by the same letter were not significantly different at the 0.10 level, as determined by Tukey's w procedure.

the drained pore space was apparently filled by aerated surface water flowing laterally through the A and O horizons toward the ditch. These results show that relatively large changes in soil morphology can occur following even small changes in hydrology.

Interpreting Morphological Features of Reduction

Morphological features of reduction simply show that the soil has been reduced at some point in its past. For example, many organic-C-based indicators show where reduction has occurred, as do redox depletions. The reduced matrix and an odor of H_2S indicate that the soil is currently reduced at the place these features are detected. On the other hand, redox concentrations indicate where oxidation has occurred in the past. By themselves, these features give no indication of how long the soils were saturated and reduced.

Occasionally, more information is desired, particularly an estimate as to whether and for how long the soils become saturated in a year of normal rainfall. Assessment of the duration of saturation is necessary for some uses, such as on-site waste disposal using septic systems. This information can be inferred by using morphological features of reduction that have been correlated to measurements of saturation.

Relating Feature Abundance to Duration of Saturation and Reduction

Morphological features that form in reduced soils range widely in their abundance. Abundance is directly related to how long the soils have been reduced, but indirectly related to how long soils have been saturated. This is because soils do not become reduced as soon as saturation begins (Table 7.2). A comparison of the abundance of redox depletions to periods of saturation and reduction is shown in Table 7.6. The data were obtained for two soils, one of which was on the backslope position and another was in the toeslope position. The amount of redox depletions varied fourfold between the two soils. The soils

TABLE 7.6

Relationship of Durations of Saturation and Fe Reduction to the Percentage of Redox Depletions in Two Soils along a Hillslope

Landscape Position[a]	Depth (cm)	Durations of Saturation (% of Year)	Durations of Fe Reduction (% of Year)	Redox Depletions (% by Weight)
Backslope	143–170	38	5	18
Toeslope	118–143	46	28	79

Source: Data from Vepraskas, M. J. and L. P. Wilding. 1983. *Soil Sci. Soc. Am. J.* 47: 1202–1208.

Note: The data show that the percentage of redox depletions is more directly related to the duration of Fe reduction than to the duration of saturation.

[a] Soil on the backslope is classified as a Plinthic Paleudult, and the soil in the toeslope position is classified as a Fragic Glossudalf.

were saturated for similar lengths of time, but reduced for longer periods in the toeslope position, which had the most redox depletions. Similar results were reported by Evans and Franzmeier (1986), Cogger and Kennedy (1992), and Couto et al. (1985), who showed that saturation by itself did not produce redox depletions.

Seasonal High Water Table Determinations

The preceding section suggests that it is impossible to develop a single relationship for all soils using the amount of gray color (redox depletions) in a soil to predict the specific length of time the soil is saturated at a given depth. An alternative approach has been to simply estimate the approximate height of the "seasonal high water table" from the presence of any redox depletions. This is the simplest way to relate redoximorphic features to saturation. Normally, it is assumed that a water table rises to the level at which redox depletions occur that have chromas of 2 or less and values of 4 or more. Note that, if the value is less than 4 and the chroma is 2 or less, then, the color is black or dark gray and is not necessarily related to saturation or reduction. The depth at which the redox depletions begin marks the level at which the seasonal high water table reaches in the soil. It is assumed that the water table will rise to this level in most years of "normal rainfall." It stays at the level of the redox depletions long enough for reduction to occur. This interpretation implies that the water table rises no farther, but this is not known unless detailed records of water table fluctuation are available. All that can be said is that the water table does not stay above the level of the low-chroma colors long enough to cause the reduction of Fe.

The advantage of using redox depletions to determine the seasonal high water table is that the determination can be made in virtually any soil, at a low cost, and without any additional information on hydrology or rainfall. The disadvantage is that there is no information on saturation frequency or duration. Nevertheless, determining the depth to the seasonal high water table in this way has proven to be useful and in general reliable for making on-site assessments as to whether a soil was suitable for septic systems. Case studies using this approach have been reported by Cogger and Kennedy (1992), Franzmeier et al. (1983), and Zobeck and Ritchie (1984a, b), among others.

New Approaches

Hydrologic models that simulate water table levels have provided another tool for developing specific relationships between soil saturation, water table fluctuation, and abundance of morphological features. Simonson and Boersma (1972) may have been the first to develop such relationships. Relationships between saturation frequency and soil color have not been developed widely, but in a few cases where it is clear that simulation modeling provides a powerful tool when making interpretations from soil morphology.

The results from Vepraskas et al. (2004) are shown in Figure 7.12. The hydrologic model used was DRAINMOD (Skaggs 1978), and daily water table levels were computed for a 32-year period using historic rainfall data and on-site calibration of the model. Figure 7.12 shows that abundance of redox depletions was related to periods of saturation lasting for 3 weeks or longer. Relationships between abundance and saturation duration changed with depth in this example. The reason for this is related to the decrease in decomposable organic materials with depth. At 90 cm, a given saturation frequency produced fewer depletions than at a depth of 30 cm. At 90 cm, roots are the major source of organic-C, and they are oriented vertically and spaced 25–50 mm apart on the outside of soil peds. Redox depletions form primarily around these widely spaced roots and occupy less volume than at 30 cm where roots are more abundant and are closely spaced.

The data in Figure 7.12 also show that the abundance of depletions at a given depth can be related to events that do not occur every year. The depletions probably occur during the wetter years and are preserved. When abundance of redox depletions falls below 2%, the events they are related to occur rarely or about once in 20 years.

The advantage of using hydrologic models to predict historic water table levels is that the data produced are very specific and allow prediction of both saturation frequency and duration. Even the occurrence of rare events can be detected. The disadvantage is that the models can be expensive to use due to their need for a variety of soil measurements. In addition, hydrologic models are generally developed for specific kinds of landscapes. For example, DRAINMOD was developed to predict how deep and far apart ditches or tile drains need to be placed in fields to lower the water table to a specific amount. It works best in level, coastal plain-type landscapes where groundwater moves laterally to streams or ditches.

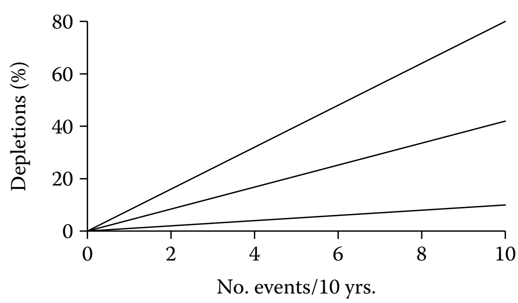


FIGURE 7.12
Relation of the frequency of saturation lasting 3 weeks or more to the percentage of redox depletions at various depths in a catena of three Ultisols in North Carolina. The saturation frequency was determined by simulating water table fluctuations over a 32-year period using the hydrologic model DRAINMOD. (From Vepraskas, M. J. et al. 2004. *Soil Sci. Soc. Am. J.* 68: 1461–1469.)

The results shown in Figure 7.12 are only for illustration because they are site specific. Different relationships will have to be developed for most other soils because the same duration of saturation will not necessarily produce the same amount of redox depletions as those shown. These differing amounts of redox depletions are caused by differences in organic matter levels, pH, temperature, and so on, which cause a given amount of saturation to produce different durations and levels of reduction.

Problem Situations

Identifying Relict Features of Reduction

A "relict feature" of reduction is one that has formed in the past and persists in the soil where it can no longer form today. Relict features of reduction make the soil appear to be wetter than it really is. They are useful in identifying soils whose hydrology has changed. A relict feature may persist in soils that were formerly saturated and reduced, but have since been historically drained by natural or artificial means such that reducing conditions no longer occur. Using relict features to detect altered hydrology can be faster than monitoring water table levels. Redoximorphic features that are either redox depletions or redox concentrations are the most likely morphological features to be relict features. The matrix must be kept reduced and can never be relict. Carbon-based organic features probably decompose too quickly for them to be preserved for more than 30 years. The single sulfur-based feature known (i.e., H_2S gas) is only found in reduced soils.

Identification of redoximorphic features that might be relict cannot be done with certainty using morphology alone, because hydrologic data are necessary to confirm that the hydrology is different than the features suggest. However, some guidelines can be given for when relict features should be suspected.

Relation to Root Channels and Cracks

In loamy and clayey soils, redoximorphic features often form around root channels or cracks, as shown in Figures 7.9 and 7.7. Redox depletions tend to form along root channels where the organic-C occurred and fueled the reduction process. These are frequently the first pores to fill with water following a heavy rain. Redox concentrations that are Fe pore linings must also occur along root channels or ped surfaces. Any time a morphological feature, which appears to have formed along a macropore, is found in the matrix or away from a pore, it can be assumed that it did not form recently and should be considered relict. Examples of this concept are schematically shown in Figure 7.13. Even features that occur in the matrix need to have a consistent relationship to the soil structure and large pores that is consistent with how they formed. For example, Fe nodules normally form in the soil matrix. If these are found in Fe depletions on ped surfaces, then, it is likely that these nodules are relict features.

Diffuse versus Sharp Boundaries

Redox concentrations form by accumulation of Fe at certain points in the soil. The amount of Fe in these features is not expected to be the same throughout the feature. Normally, Fe

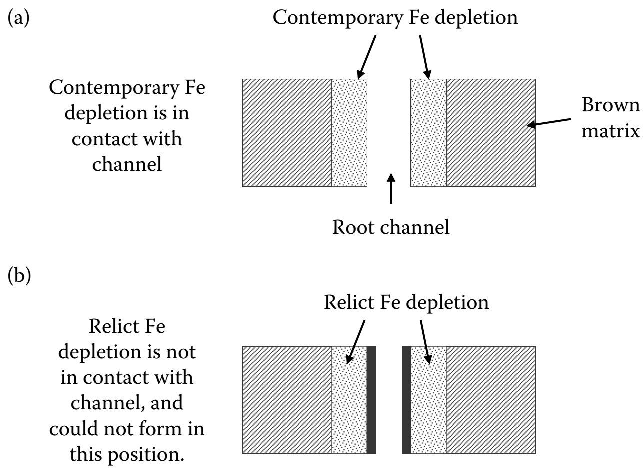


FIGURE 7.13
Relationship between the location of a redox depletion to a root channel in a soil where the depletion has formed recently and one where it is considered relict. Where the features are thought to be contemporary (a), the depletion abuts the channel, which is the position required for it to form by the process shown in Figure 7.9. In the second case (b), the depletion is separated from the channel by an Fe-rich clay coating. This coating suggests the depletion had to form before the clay was deposited; otherwise, the Fe in the clay coating would have been reduced and removed from the coating.

quantities decrease from the center of the Fe concentration toward the soil matrix. The zone of decreasing Fe concentration is frequently described as a *diffuse boundary* (see Figure 7.2). It is sometimes seen as a ring or halo around the Fe concentration that has a slightly different color than the main part of the Fe concentration itself. Diffuse boundaries are assumed to indicate that the feature is forming or has formed in the recent past. In other words, it is reflecting the current hydrologic conditions.

When redox concentrations begin to dissolve, or are mixed into the matrix, they acquire *sharp boundaries* with the matrix. In this case, the features are no longer forming and are relict. If virtually all the redox concentrations in a horizon have such sharp boundaries, then, it is likely that the hydrology has changed to make the soils oxidized year round. However, the underlying horizons should also be examined to find features that may be forming and to determine the exact appearance of Fe concentrations with diffuse boundaries in that soil.

When No Indicators Are Present

Occasionally, soils that are suspected of being seasonally saturated and reduced do not show the common morphological features indicative of reduced soils. The reasons morphological features of reduction do not form are not completely understood but probably relate to the fact that little Fe reduction occurs or insufficient time has elapsed to form redoximorphic features under the specific conditions. Iron reduction is also limited by the soil having low amounts of organic-C at the time of saturation, a high pH, which makes Fe reduction occur only at very low Eh values as discussed in Chapter 4, high levels of Mn oxides in the soil, or large amounts of dissolved O_2 in the water. Identifying hydric soils in areas where no morphological indicators of reduction can be found requires direct measurements of saturation and reduction. Reducing conditions will have to be documented using dyes that react with Fe^{2+}, redox electrodes, or an indicator of reduction in soil tubes (Childs 1981; Chapter 4).

Problematic hydric soils often occur on floodplains (Lindbo 1997). These relatively young geomorphic surfaces are particularly difficult areas to identify hydric soils due to the frequent deposition of material on the surface, in essence, restarting the soil formation clock. Redox features that do form in these settings are often less pronounced and fewer in quantity. Dark, organic-rich hydric soils are also complicated to identify due to the masking qualities of the organic materials. Mollisols with their low value and chroma often lack the necessary quantity and distinctness in redox features that are required to meet a field indicator of hydric soil (Thompson and Bell 1998; USDA NRCS 2010). Vernal pools containing salt-affected soils and other soils with high pH often lack redox features that would be found in soils with a similar hydrology (Clausnitzer et al. 2003; O'Geen et al. 2008). As soil pH increases, the redox potential required to reduce redox-sensitive species decreases and as a result, becomes more difficult. Soils formed in red parent materials inherit high concentrations of Fe that result in poor redox feature formation. Specialized field indicators and other techniques have been developed to help identify hydric soils in these parent materials, landscapes, and soil types (USDA NRCS 2010; Chapter 8).

False Redoximorphic Features

Gray Parent Materials

Some soil parent materials have virtually no Fe minerals coating the particle surfaces and contain no Fe-bearing minerals. These materials have a gray color and will remain gray regardless of whether the soils that develop in them become reduced or not. Soils that develop in these deposits will have an A horizon that formed by the accumulation of organic debris and a C horizon. Such soils can be well drained, but because of their gray color, which resembles an Fe depletion, they will have the appearance of being seasonally saturated and reduced. This condition should be suspected whenever the parent material (C horizon) is gray in color due to a naturally low amount of Fe. Soils whose parent materials consisted of gray sands will remain gray, and no Fe-based morphological features of reduction will develop. The only morphological indicators of reduction that will develop are organic-C-based features or S-based features. Landscape position should also be examined for signs of it being where seasonally saturated soils would be expected. Such positions include the base of steep slopes and concave positions on flat or gently sloping surfaces.

E Horizons

E horizons are layers in the subsoil that developed a gray color through soil-forming processes that may or may not include Fe reduction. E horizons form by eluviation or movement of Fe, clay, and organic matter out of the soil layer resulting in a gray, leached horizon. They usually occur below A horizons and must overlie zones of accumulation, such as Bt horizons. E horizons frequently have a chroma of 2 or less and a value of 4 or more when the sand and silt grains have been stripped of Fe oxide coatings. The loss of Fe can occur by reduction processes (as in hydric soils), or it can occur because organic acids produced in A or O horizons leach into E horizons and dissolve Fe^{3+} minerals off particle surfaces.

In the latter case, the soils are not considered to be hydric. The E horizon is similar to an A horizon in texture and chemical composition except that it does not contain organic matter, Fe, or clay in the amounts found in the A horizon.

Grayish-colored E horizons that formed in reduced soils are identified by redox concentrations in an abundance that usually exceeds 2%. The concentrations show that Fe has been reduced, moved, and reoxidized into Fe masses or pore linings. In addition, the horizons below the E should also be examined for morphological features of reduction. When redoximorphic features occur both within and below the E horizon, it is likely that the E horizon was formed under reducing conditions.

Geological Materials below the Rooting Zone

Morphological features of reduction need a source of organic carbon to form. The carbon is in greatest concentration near the surface and decreases in concentration with depth. Below a depth of approximately 1 m, the organic carbon is usually found around roots, and it becomes more scarce with increasing depth. C horizons below 1 m can contain features that are, or appear to be, redoximorphic features. While the features may have formed by oxidation–reduction processes, they must be interpreted cautiously because it is not always clear when the features formed. For example, Schoeneberger et al. (1992) described redoximorphic features along fractures in saprolite. Similar features could also be formed by hydrothermal fluids moving upward to the C horizon (see Figures 11–33 in Guilbert and Park 1986). Heated hydrothermal fluids can become strongly reducing when they pass through layers of graphite (Guilbert and Park 1986). The passage through the carbon-rich graphite produces solutions capable of reducing Fe. While such formation may be rare on a global scale, it does happen. When such features are formed deeply in the soil, they can probably be preserved for thousands of years.

Summary

Morphological features of reduced soils form by oxidation–reduction reactions that occur when the soils are anaerobic. The reduction of O_2, Mn^{4+} and Fe^{3+} minerals, and SO_4^{2-} are responsible for the formation of most of the morphological features of reduction. When O_2 is reduced, organic matter accumulation exceeds decomposition, and this leads to the development of features rich in organic-C, such as layers of peat or muck. When Fe or Mn minerals are reduced, redoximorphic features develop. The reduction of SO_4^{2-} leads to the production of H_2S gas. Morphological features of reduction only show that a soil was reduced at some point in its history. Their abundance can be related to the frequency of saturation, but such relationships are expected to be site specific and are not widely understood at this time. Hydrologic alterations caused by soil drainage or wetland construction produce changes in morphological features of reduction that can be detected in soils within 3 years or less of wetland construction and in less than 30 years following ditching. Such rates of formation will vary by region due to differences in temperature, organic carbon levels, as well as other factors. Relict features are those that occur in soils that have been historically drained, but which formed under wetter conditions. They may be identified by their relationship to root channels, cracks, and lack of diffuse boundaries. False redoximorphic features also occur in soils with gray E horizons or gray parent materials. These do not develop under reducing conditions and cannot be used to identify hydric soils.

References

Ahmad, N. R. and P. H. Nye. 1990. Coupled diffusion and oxidation of ferrous iron in soils. I. Kinetics of oxygenation of ferrous iron in soils. *J. Soil Sci.* 41: 395–409.

Bell, J. C. and J. L. Richardson. 1997. Aquic conditions and hydric soil indicators for Aquolls and Albolls. In M. J. Vepraskas, and S. W. Sprecher (Eds.) *Aquic Conditions and Hydric Soils: The Problem Soils.* SSSA Special Publication No. 50. Soil Science Society of America, Inc., Madison, WI, pp. 23–40.

Bidwell, O. W., D. A. Gier, and J. C. Cipra. 1968. Ferromagnesian pedotubules on roots of *Bromus inermis* and *Andropogon gerardii. International Congress of Soil Science Transactions.* Adelaide, Australia, IV, pp. 683–692.

Brewer, R. 1964. *Fabric and Mineral Analysis of Soils.* John Wiley & Sons, New York.

Chen, C. C., J. B. Dixon, and F. T. Turner. 1980. Iron coatings on rice roots: Morphology and models of development. *Soil Sci. Soc. Am. J.* 44: 1113–1119.

Childs, C. W. 1981. Field test for ferrous iron and ferric–organic complexes on exchange sites (in water soluble forms) in soils. *Aust. J. Soil Res.* 19: 175–180.

Clausnitzer, D., J. H. Huddleston, E. Horn, M. Keller, and C. Leet. 2003. Hydric soils in a Southeastern Oregon vernal pool. *Soil Sci. Soc. Am. J.* 67: 951–960.

Cogger, C. G. and P. E. Kennedy. 1992. Seasonally saturated soils in the Puget Lowland. I. Saturation, reduction, and color patterns. *Soil Sci.* 153(6): 421–433.

Couto, W., C. Sanzonowicz, and A. O. de Barcellos. 1985. Factors affecting oxidation–reduction processes in an Oxisol with a seasonal water table. *Soil Sci. Soc. Am. J.* 49: 1245–1248.

Daniels, R. B., E. E. Gamble, and L. A. Nelson. 1971. Relations between soil morphology and water-table levels on a dissected North Carolina Coastal Plain surface. *Soil Sci. Soc. Am. Proc.* 35: 781–784.

Environmental Laboratory. 1987. *Corps of Engineers Wetlands Delineation Manual.* Technical Report Y-87-1. U.S. Army Engineers Waterways Experiment Station, Vicksburg, MS.

Evans, C. V. and D. P. Franzmeier. 1986. Saturation, aeration, and color patterns in a toposequence of soils in north-central Indiana. *Soil Sci. Soc. Am. J.* 50: 975–980.

Fiedler, S. and M. Sommer. 2004. Water and redox conditions in wetland soils—Their influence on pedogenic oxides and morphology. *Soil Sci. Soc. Am. J.* 68: 326–335.

Franzmeier, D. P., J. E. Yahner, G. C. Steinhardt, and H. R. Sinclair. 1983. Color patterns and water table levels in some Indiana soils. *Soil Sci. Soc. Am. J.* 47: 1196–1202.

Gallaher, R. N., H. F. Perkins, and K. H. Tan. 1974. Classification, composition, and mineralogy of iron glaebules in a Southern Coastal Plain soil. *Soil Sci.* 117: 155–164.

Genthner, M. H., W. L. Daniels, R. L. Hodges, and P. J. Thomas. 1998. Redoximorphic features and seasonal water table relations, upper coastal plain Virginia. In M. C. Rabenhorst, and P. A. McDaniel (Eds.) *Quantifying Soil Hydromorphology.* SSSA Special Publication 54. SSSA, Madison, WI, pp. 43–60.

Gilman, K. 1994. *Hydrology and Wetland Conservation.* John Wiley & Sons, New York.

Griffin, R. 2008. *Use of Alpha-Alpha-Dipyridyl.* Hydric Soils Technical Note 8, National Technical Committee for Hydric Soils. USDA, Natural Resources Conservation Service.

Guilbert, J. M. and C. F. Park Jr. 1986. *The Geology of Ore Deposits.* W.H. Freeman and Company, New York.

Hayes Jr., W. A. and M. J. Vepraskas. 2000. Morphological changes in soils produced when hydrology is altered by ditching. *Soil Sci. Soc. Am. J.* 64: 1893–1904.

Hudnall, W. H., A. Szogi, B. A. Touchet, J. Diagle, J. P. Edwards, and W. C. Lynn. 1990. *Guidebook for Louisiana. VIII International Soil Correlation Meeting: Classification and Management of Wet Soils*. Agronomy Department, Louisiana State University, Baton Rouge.

Jackson, M. L. 1969. *Soil Chemical Analysis—Advanced Course*, 2nd ed. 10th printing. Published by the author, Madison, WI.

Jacobs, P. M., L. T. West, and J. N. Shaw. 2002. Redoximorphic features as indicators of seasonal saturation, Lowndes County, Georgia. *Soil Sci. Soc. Am. J.* 66: 315–323.

Lindbo, D. L. 1997. Entisols: Fluvents and fluvaquents: Problems recognizing aquic and hydric conditions in young, flood plain soils. In M. J. Vepraskas, and S. W. Sprecher (Eds.) *Aquic Conditions and Hydric Soils: The Problem Soils*. SSSA Special Publication No. 50, Soil Science Society of America, Madison, WI, pp. 133–151.

McBride, M. B. 1994. *Environmental Chemistry of Soils.* Oxford University Press, New York.

Meek, B. D., A. J. McKenzie, and L. B. Gross. 1968. Effects of organic matter, flooding time and temperature on the dissolution of iron and manganese from soil *in situ*. *Soil Sci. Soc. Am. Proc.* 32: 634–638.

Megonigal, J. P., S. P. Faulkner, and W. H. Patrick Jr. 1996. The microbial activity season in southeastern hydric soils. *Soil Sci. Soc. Am. J.* 60: 1263–1266.

Megonigal, J. P., W. H. Patrick, and S. P. Faulkner. 1993. Wetland identification in seasonally flooded forest soils: Soil morphology and redox dynamics. *Soil Sci. Soc. Am. J.* 57: 140–149.

Mendelssohn, I. A., B. A. Kleiss, and J. A. Wakeley. 1995. Factors controlling the formation of oxidized root channels: A review. *Wetlands* 15(1): 37–46.

Morgan, C. P. and M. H. Stolt. 2006. Soil morphology–water table cumulative duration relationships in Southern New England. *Soil Sci. Soc. Am. J.* 70: 816–824.

O'Geen, A. T., W. A. Hobson, R. A. Dahlgren, and D. B. Kelley. 2008. Evaluation of soil properties and hydric soil indicators for vernal pool catenas in California. *Soil Sci. Soc. Am. J.* 72: 727–740.

Ponnamperuma, F. N. 1972. The chemistry of submerged soils. *Adv. Agron.* 24: 29–96.

Rabenhorst, M. C. 2005. Biologic zero: A soil temperature concept. *Wetlands* 25: 616–621.

Rhoton, F. E., J. M. Bigham, and D. G. Schultze. 1993. Properties of iron–manganese nodules from a sequence of eroded fragipan soils. *Soil Sci. Soc. Am. J.* 57: 1386–1392.

Russell, R. S. 1977. *Plant Root Systems.* McGraw-Hill Book Company, London.

Schoeneberger, P. J., S. B. Weed, A. Amoozegar, and S. W. Buol. 1992. Color zonation associated with fractures in a felsic gneiss saprolite. *Soil Sci. Soc. Am. J.* 56: 1855–1859.

Schwertmann, U. and R. M. Taylor. 1989. Iron oxides. In J. B. Dixon, and S. W. Weed (Eds.) *Minerals in Soil Environments*, 2nd ed. Soil Science Society of America, Madison, WI, pp. 379–438.

Simonson, G. H. and L. Boersma. 1972. Soil morphology and water table relations: II. Correlation between annual water table fluctuations and profile features. *Soil Sci. Soc. Am. J.* 36: 649–653.

Skaggs, R. W. 1978. *A Water Management Model for Shallow Water Table Soils.* Technical Report 134. Water Research Institute, Raleigh, NC.

Soil Survey Staff. 2010. *Keys to Soil Taxonomy*, 11th ed. USDA, National Research Conservation Service, Washington, DC.

Szogi, A. A. and W. H. Hudnall. 1998. Soil morphology and frequency of diagnostic wet soil conditions. In M. C. Rabenhorst, and P. A. McDaniel (Eds.) *Quantifying Soil Hydromorphology*. SSSA Special Publication 54. SSSA, Madison, WI, pp. 61–76.

Thompson, J. A. and J. C. Bell. 1998. Hydric conditions and hydromorphic properties within a Mollisol catena in Southeastern Minnesota. *Soil Sci. Soc. Am. J.* 62: 1116–1125.

USDA NRCS. 2010. *Field Indicators of Hydric Soils in the United States*. Version 7.0. In L. M. Vasilas, G. W. Hurt, and C. V. Noble (Eds.) USDA, NRCS, in cooperation with the National Technical Committee for Hydric Soils.

Veneman, P. L. M., M. J. Vepraskas, and J. Bouma. 1976. The physical significance of soil mottling in a Wisconsin toposequence. *Geoderma* 15: 103–118.

Vepraskas, M. J. 1996. *Redoximorphic Features for Identifying Aquic Conditions.* Technical Bulletin 301. North Carolina Agricultural Experiment Station, Raleigh, NC.

Vepraskas, M. J. and J. Bouma. 1976. Model experiments on mottle formation simulating field conditions. *Geoderma* 15: 217–230.

Vepraskas, M. J. and W. R. Guertal. 1992. Morphological indicators of soil wetness. In J. M. Kimble (Ed.) *Proceedings of the Eighth International Soil Correlation Meeting* (*VIII ISCOM*): *Characterization, Classification, and Utilization of Wet Soils.* USDA Soil Conservation Service, National Soil Survey Center, Lincoln, NE, pp. 307–312.

Vepraskas, M. J., X. He, D. L. Lindbo, and R. W. Skaggs. 2004. Predicting soil saturation frequency and duration using soil color in two toposequences. *Soil Sci. Soc. Am. J.* 68: 1461–1469.

Vepraskas, M. J., J. L. Richardson, and J. P. Tandarich. 2006. Dynamics of redoximorphic feature formation under controlled ponding in a created riverine wetland. *Wetlands* 26: 486–496.

Vepraskas, M. J., J. L. Richardson, J. P. Tandarich, and S. J. Teets. 1999. Dynamics of hydric soil formation across the edge of a created deep marsh. *Wetlands* 19(1): 78–89.

Vepraskas, M. J. and L. P. Wilding. 1983. Albic neoskeletans in argillic horizons as indices of seasonal saturation and iron reduction. *Soil Sci. Soc. Am. J.* 47: 1202–1208.

West, L. T., J. N. Shaw, E. R. Blood, and L. K. Kirkman. 1998. Correlation of water tables to redoximorphic features in the Daugherty Plain, southwest Georgia. In M. C. Rabenhorst, and P. A. McDaniel (Eds.) *Quantifying Soil Hydromorphology.* SSSA Special Publication 54. SSSA, Madison, WI, pp. 247–258.

Zobeck, T. M. and A. Ritchie. 1984a. Analysis of long-term water table depth records from a hydrosequence of soils in central Ohio. *Soil Sci. Soc. Am. J.* 48: 119–125.

Zobeck, T. M. and A. Ritchie Jr. 1984b. Relation of water table depth and soil morphology in two clay-rich soils of Northwestern Ohio. *Ohio J. Sci.* 84(5): 228–236.

8

Identifying Hydric Soils in the Landscape

Lenore M. Vasilas and Jacob F. Berkowitz

CONTENTS

Introduction

The National Technical Committee for Hydric Soils (NTCHS) defines a hydric soil as "… a soil that formed under conditions of saturation, flooding, or ponding long enough during the growing season to develop anaerobic conditions in the upper part" (Federal register 1994, pp. 94–16835). The term "hydric soil" was created for the purposes of identifying soils found in wetlands (Vepraskas and Sprecher 1997). The identification of hydric soils and their boundaries is most significant for the purposes of identifying areas that are protected under Section 404 of the Clean Water Act and/or the Swampbuster Provision of the Food Security Act (Tiner 1999). All delineation methodologies for the purposes of identifying soils meeting the definition of a hydric soil are deliberated and approved by the NTCHS and can be found in the *1987 Corps of Engineers Delineation Manual* and associated regional supplements and on the NTCHS website at http://www.nrcs.usda.gov/wps/portal/nrcs/main/soils/use/hydric/ (Environmental Laboratory 1987; USACE 2012).

Typically, hydric soils occur in landscape positions where water accumulates (e.g., floodplains, depressions) and field investigations focus in these areas. Hydric soils exhibit readily identifiable morphological patterns, including dark soils high in organic matter content and soils containing redoximorphic features. On-site hydric soils identification requires the completion of a hydric soils data form, including measurements describing the depth of soil layers, soil colors, redoximorphic features, and soil texture. The completed hydric soil data form allows for the determination of the presence or absence of hydric soils.

In this chapter, we discuss methods for identifying soils that meet the hydric soil definition. These include: (1) off-site methods capable of locating areas that likely contain hydric soils and (2) on-site field data collection techniques for recognizing and describing morphological features that identify and delineate the boundary of hydric soils. Additionally, we introduce terminology and concepts that aid in the identification of hydric soils.

Off-Site Investigations for the Presence of Hydric Soils

Web Soil Survey and the National List of Hydric Soils

The Natural Resources Conservation Service (NRCS) publishes regional soil surveys that contain the most comprehensive information for locating hydric soils (USDA NRCS 1993). The NRCS maintains the official soil surveys on Web Soil Survey (WSS), available at http://websoilsurvey.sc.egov.usda.gov/App/HomePage.htm WSS uses information on soil properties from the NRCS database to generate hydric soils lists and interpretive maps identifying areas that likely contain hydric soils (USDA 1985). The NTCHS also creates a National List of Hydric Soils (http://www.nrcs.usda.gov/wps/portal/nrcs/main/soils/use/hydric/) representing a yearly compilation of all map units containing hydric soils throughout the United States.

The National List of Hydric Soils utilizes four criteria evaluating soil map unit components to determine if they classify as hydric soils in the NRCS database. The criteria (NTCHS 2012) are as follows:

1. All Histels except Folistels and Histosols except Folists; or
2. Map unit components in Aquic suborders, great groups, or subgroups, Albolls suborder, Historthels great group, Histoturbels great group, or Andic, Cumulic, Pachic, or Vitrandic subgroups that
   a. On the basis of the range of characteristics for the soil series, will at least in part meet one or more *Field Indicators of Hydric Soils in the United States*, or
   b. Show evidence that the soil meets the definition of a hydric soil
3. Map unit components that are frequently ponded for a long duration or a very long duration during the growing season that
   a. On the basis of the range of characteristics for the soil series, will at least in part meet one or more *Field Indicators of Hydric Soils in the United States*, or
   b. Show evidence that the soil meets the definition of a hydric soil; or
4. Map unit components that are frequently flooded for a long duration or a very long duration during the growing season that
   a. On the basis of the range of characteristics for the soil series, will at least in part meet one or more *Field Indicators of Hydric Soils in the United States*, or
   b. Show evidence that the soils meet the definition of a hydric soil

If a map unit component meets any of the criteria, then, that component classifies as hydric in the NRCS soils database and all map units that contain that component will be identified as containing hydric soils on the National List of Hydric Soils and other hydric soils lists (Figure 8.1).

Use of WSS

WSS allows you to produce interpretive tables and maps for an area based on soil characteristics. The WSS provides valuable off-site information concerning the likelihood that a hydric soil exists in your area of interest. Within WSS, the Hydric Rating by Map Unit represents the most valuable tool for identifying areas that likely contain hydric soil (Figure 8.2). This interpretation categorizes map units based on the percentage of that map unit considered a hydric soil. The interpretive map uses a five-category system of hydric (100%), predominately hydric (66%–99%), partially hydric (33%–65%), predominately nonhydric (1%–32%), and nonhydric (0%). The WSS also produces several other soil reports on hydric soils, including a list of all map unit components in your area of interest that classify as hydric and their percentages.

Field Sampling for the Identification of Hydric Soils

Introduction

While WSS and other off-site tools aid in the identification of areas potentially containing hydric soils, most efforts require field sampling to ensure an area meets the hydric soil definition. Field sampling also allows for the delineation of the boundary between hydric soils and nonhydric soils (Environmental Laboratory 1987). Field sampling requires that

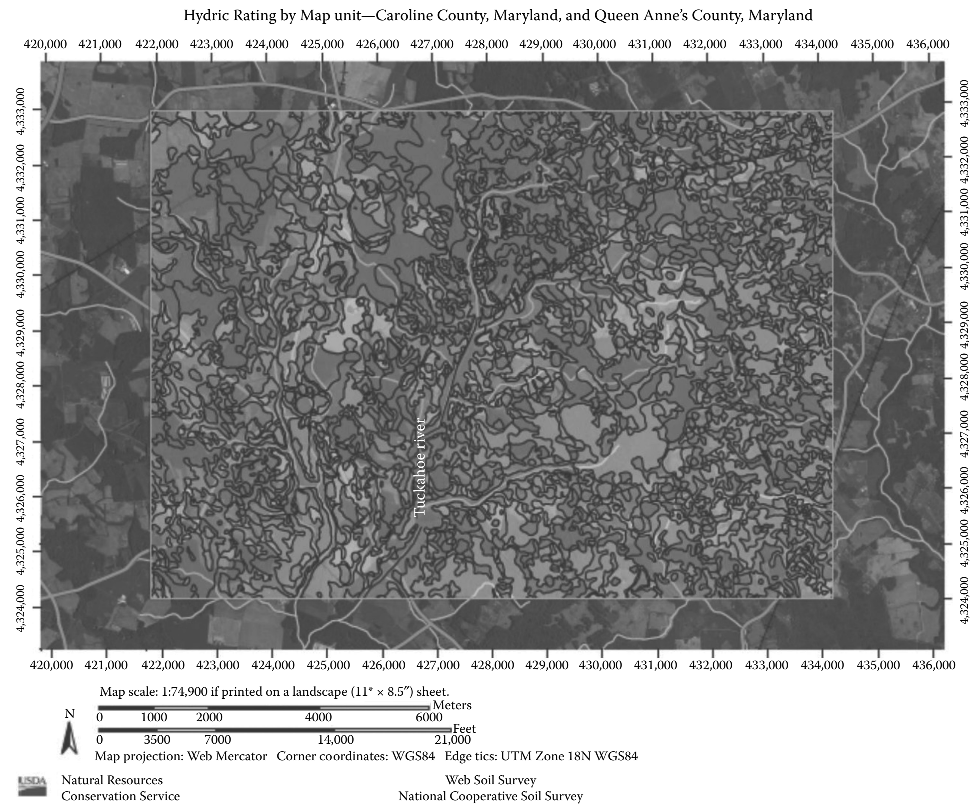


FIGURE 8.1
Grayscale depiction of a map generated using Web Soil Survey (WSS) depicting Hydric Rating by Map Units at a large scales. The WSS generates color maps, making the categories easy to distinguish between soil map units. Note that field investigations of hydric soils should focus on areas containing hydric and/or partially hydric soil map units.

you (1) identify areas in the landscape that likely contain hydric soils, (2) excavate a soil pit to 25 cm or deeper, (3) describe characteristics in the soil that help identify hydric soil morphologies, and (4) compare the soil description to a list of *Field Indicators of Hydric Soils in the United States* (USDA NRCS 2010; USACE 2012). This approach identifies soils that meet the definition of a hydric soil. This chapter addresses numbers one through three above, focusing on sampling and data collection techniques; the subsequent chapters discuss the application of hydric soils indicators. In this chapter, we introduce the data required to complete a hydric soils data form (Figure 8.3). Required measurements include the depth of each soil layer, determinations of the matrix color and abundance, redoximorphic redox color and abundance, redoximorphic feature type and location, and the soil texture.

The identification and delineation of hydric soils requires field sampling because soil mapping and other off-site resources occur at a variety of scales, many of which remain too large to determine the occurrence and extent of hydric soils within a particular area of interest (Tiner 1999). Fortunately, biogeochemical processes occurring in wetlands and saturated soils result in soil characteristics that can be seen, felt, or smelled during most on-site field visits (USDA NRCS 2010). As discussed elsewhere in this chapter and within the

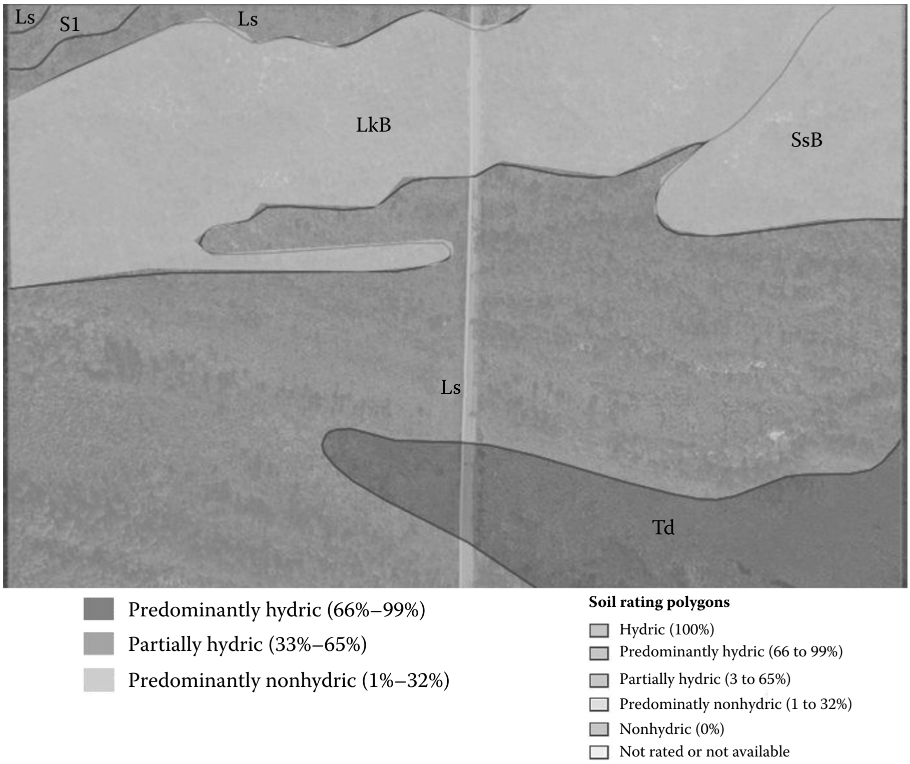


FIGURE 8.2
Grayscale maps generated by Web Soil Survey (WSS) depicting Hydric Rating by Map Unit at a fine scale. The WSS generates color maps, making the categories easy to distinguish between soil map units. Note that hydric soils are more likely to occur in the darker shaded portions of and field investigations for hydric soils should focus in those areas.

Profile Description: (Describe to the depth needed to document the indicator or confirm the absence of indicators.)

Depth (cm)	Matrix Color (moist)	Matrix %	Redox Features Color (moist)	Redox Features %	Type[1]	Loc[2]	Texture	Remarks

[1]Type: C=Concentration, D=Depletion, RM=Reduced Matrix, CS=Covered or Coated Sand Grains. [2]Location: PL=Pore Lining, M=Matrix.

FIGURE 8.3
The data form used in the identification of hydric soils. *Note*: The major elements listed at the top of the form. Required measurements include: the depth of each soil layer, determinations of the matrix color and abundance, redoximorphic redox color and abundance, redoximorphic feature type and location, and the soil texture. From USACE. 2012. In J. S. Wakeley, R. W. Lichvar, C. V. Noble, and J. F. Berkowitz (Eds.) *Regional Supplement to the Corps of Engineers Wetland Delineation Manual: Northcentral and Northeast Region (Version 2.0)*. ERDC/EL TR-08-28. U.S. Army Engineer Research and Development Center, Vicksburg, MS.

scientific literature, extended periods of soil saturation result in the anaerobic conditions and chemical reduction of electron potentials (Reddy and DeLaune 2008). These factors cause a number of changes in soils that prove useful for hydric soil identification (Fanning and Fanning 1989; Richardson and Daniels 1993). Notably, biogeochemical changes associated with hydric soils result in three main identifiable characteristics including:

1. A decrease in organic carbon decomposition rates and the accumulation of organic matter in hydric soils. Accumulation of organic matter causes many hydric soils to exhibit dark colors, a persistence of undecomposed vegetative plant material in the soil, and soils that often feel slippery or greasy (Sahrawat 2003).
2. The translocation of iron and manganese compounds results in the formation of distinct color patterns in hydric soils. The movement of iron and manganese yields areas of gray depletions where iron and manganese have been removed. Additionally, the accumulation of iron oxides along pore linings, soil cracks, and in the soil matrix results in yellow-brown to reddish iron concentrations. When present, manganese oxides accumulate in areas exhibiting bluish-black concentrations (Birkeland 1999; Vepraskas 2004).
3. The reduction of sulfate compounds in soils experiencing extended periods of saturation, flooding, or ponding results in the formation of hydrogen sulfide gas, producing a distinct "rotten egg odor" characteristic of many hydric soils (Faulkner 2004).

The distinct morphologies listed above make the field identification and delineation of hydric soils approachable and rapid for the nonsoil scientist. However, the documentation of soil morphologies observed in the field requires several steps. The sections below discuss each of the required steps, providing guidance concerning site selection, field-sampling techniques, and data collection for the purposes of hydric soil delineation.

Site Selection

Sampling should focus on positions in the landscape where hydric soils (and wetlands) most likely occur. Hydric soils form in areas characterized by frequent, seasonal, or periodic flooding, ponding, or saturation sufficient to meet the hydric soil definition. These conditions exist in a variety of landscape positions and geomorphologies including tidal areas, riparian areas, depressional areas, extensive flats, concave slopes, fringes of aquatic environments, and areas receiving overland or groundwater discharges associated with the base of a slope (Daniels et al. 1971; National Research Council 1995). WSS and the Hydric Rating by Map Unit map provide information that can guide you to areas that likely contain hydric soils and help you determine where field data collection should occur. Other information such as wetness signatures on aerial photography (USDA NRCS 1997), vegetative community breaks (Cox and Moore 1993), changes in local topography (Mitsch and Gosselink 2007; USDA NRCS 2010), and signs of wetland hydrology (Tiner 1999; USACE 2012) also help to determine where sampling should take place. Once the approximate hydric/nonhydric boundary is identified, soil pits are excavated to examine soil morphology and determine if hydric soils occur. Data should be collected in both the hydric soil and the nonhydric soil documenting that the assessment of the hydric soil boundary is correct.

Items Required for Describing Hydric Soils

The items required for collecting soil data in support of a hydric soil determination or delineation include a base map (preferably an aerial photograph or copy of the area produced using WSS); hydric soil data form (Figure 8.3); spade or sharp shooter; measuring stick or tape; knife, screw driver, or other tool for cleaning the face of the soil; nails or golf tees for marking horizon breaks; Munsell® Soil Color Chart (Gretag/Macbeth 2000); a 10× magnitude hand lens; water bottle or spray bottle of water; a copy of the *Field Indicators of Hydric Soils in the United States* (USDA NRCS 2010); regional supplement to the *Corps of Engineers Wetland Delineation Manual* (USACE 2012) for the area of interest; or a locally prepared list of field indicators of hydric soils.

Other helpful items include a cloth or tarp for laying out the soil: a bucket auger or probe for identifying the boundary between hydric soils and nonhydric soils or if you need to observe the soil below 50 cm; α-α dipyridyl dye (NTCHS 2009); and 2% hydrogen peroxide solution (Vasilas and Vasilas 2004).

Digging a Soil Pit

A detailed observation and description of the upper 25–50 cm of the soil is essential when identifying hydric soils. In most situations, observation of the upper 25–50 cm proves sufficient for making hydric soil determinations. A spade slice of the soil can be used for examinations of the upper part of the soil (Figure 8.4). However, some cases necessitate observations below 25–50 cm. A bucket auger is a sufficient tool for excavation deeper than 50 cm.

Collecting accurate hydric soil data requires the removal of an intact spade slice of soil from the side of a soil pit. The spade slice will then be described and the data recorded on the hydric soil data form (Figure 8.3). To accomplish this, excavate the upper part of the soil by digging a pit approximately 50 × 50 × 50 cm. Make a cut on either side of the soil slice with the spade, cutting any roots causing difficulty in excavation. Insert the spade into the side of the soil pit creating a soil slice at least 15 cm thick and about 50 cm deep. Carefully excavate the soil slice from the pit by slowly tipping the spade backward until the soil slice can be removed. Lay the slice on the ground or a tarp or cloth so that the soil can be described (Figure 8.4). Ensure that the soil stays together so that accurate measurements can be collected. The length of the soil slice should match the depth of the pit. If a soil slice cannot be removed for any reason (i.e., too many roots, too sandy, and to gravelly), describe the soil by expanding the size of the pit and examining the soil exposed on the side of the pit (Figure 8.4).

Pick the face of the slice with a knife, screw driver, or other tool exposing a natural face void of smearing caused by the spade. If using the alternative technique of describing the side of the hole, then, pick the face of the pit with a knife, screw driver, or other tool to remove the smear of the spade.

Use a bucket auger for observations made below the depth of your spade slice. When using the bucket auger, only turn the auger about six times. This limits soil compaction in the bucket. Once you remove a bucket of soil, lay the sample at the bottom of your spade slice. Take a measurement of the depth in the pit and then measure the length of the laid-out soil to make sure the measurements are the same. If they differ, adjust the excavated soil so that the measurement corresponds with the depth of the pit.

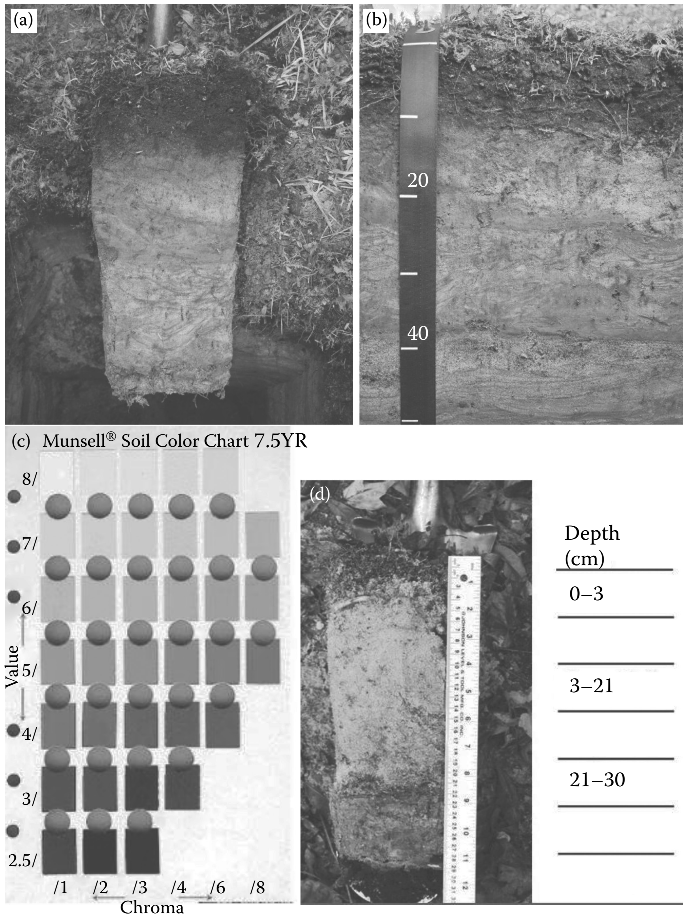


FIGURE 8.4
Clockwise from top left: (a) intact soil spade slice removed from the soil pit, (b) soil profile examining the side of the soil pit, (c) soil slice displaying three distinct layers, and (d) a grey-scale image of a page from the Munsell® Soil Color Chart. Note the hue for this page (7.5YR) located in the upper right corner of the page. Value is found along the left hand side of the page and ranges from 2.5 (darker colors) to 8 (lighter colors). Chroma is located along the bottom of the page and ranges from 1 to 8.

Writing a Soil Description for the Identification of Hydric Soils

Selecting and Measuring Soil Layers

Observe the excavated soil slice from top to bottom. Look closely for observable changes in soil characteristics such as color, texture, organic matter content, redoximorphic feature abundance, structure, or root distribution (USDA NRCS 2010). Using golf tees, nails, or other markers, separate the soil into layers based on the observed changes. These layers are called soil horizons (Singer and Munns 2002). The identification of hydric soils requires the description of features located within each soil layer or horizon (USDA NRCS 2010).

Once horizons have been identified, make measurements and record the depth of each layer on the data form (Figures 8.3 and 8.4). Measurements start with zero at the soil surface. The measurement of the deepest horizon examined should match the depth at the bottom of the soil pit.

Describing Soil Color

Soil color remains one of the most obvious and commonly reported soil characteristics (Soil Survey Staff 1951; Post et al. 1993). Soil color descriptions utilize the Munsell Soil Color Chart that applies three parameters (i.e., hue, value, and chroma) in the characterization of soil colors (Figure 8.4; Gretag/Macbeth 2000; Chapter 1). When recording soil colors on the data form, first record the hue and then record the value over the chroma as in a fraction (i.e., hue value/chroma). We provide a brief discussion of soil color components and color patterns in soils below. See the introductory material located in the front of the Munsell Soil Color Chart as well as Bigham and Ciolkosz (1993) for additional information regarding soil color determinations.

Aspects of Soil Color

1. Hue

   Hue represents the spectral color exhibited by a soil. The Munsell Soil Color Chart lists hues in the upper right-hand corner of most pages (Figure 8.4). However, the gley pages (i.e., GLEY1 and GLEY 2) list the hue along the bottom of the page. Each hue combines numbers and letters indicating the distribution of colors present. Letter designations include red (R), yellow (Y), neutral (N), green (G), blue (B), and purple (P). Hues either utilize one color (e.g., R) or a combination of different colors. For example, the hue designation YR refers to the yellow-red colors. Number designations indicate the purity of color, with 5 representing a pure color. For example, a hue of 5Y denotes a pure yellow. Once a soil color is matched to the most appropriate color chip, record the hue including both the numbers and letter (e.g., 10YR, 2.5Y) on the data form. If a soil color occurs between two hues, round it to the closest matching hue.

2. Value

   Value represents the lightness or darkness of a soil color. The Munsell Soil Color Chart lists values vertically along the left-hand side of the page (Figure 8.4). Values typically range from 2 or 2.5 to 8. Lower values are assigned to dark soil colors and higher values describe light or white soil colors. Therefore, a low value generally indicates dark soil colors associated with high organic matter content. If a soil color occurs between two values, round it to the closest matching value.

3. Chroma

   Chroma represents the saturation or intensity of a soil color. The Munsell Soil Color Chart lists chromas horizontally across the bottom of most pages (Figure 8.4). See the Munsell® Color Name Diagram for information regarding gley-page chroma designations (Gretag/Macbeth 2000). Unlike hue and value designations, never round soil chroma measurements. If the soil color appears between two color chips, record this on the data form. For example, if a soil color is between 7.5YR 4/2 and 7.5YR 4/3, record the color as 7.5YR 4/2+ or estimate the actual chroma (e.g., 7.5YR 4/2.5).

All soil colors descriptions should utilize moist soils. Break a clod of soil, exposing the interior of the soil ped. If the soil appears dry, moisten it with a spray bottle of water. If the soil is very wet or saturated, note this on the data sheet. Do not allow very wet or saturated soils to dry before assessing color as the color may change as the iron oxidizes. Hold the Munsell Soil Color Chart with natural sunlight coming over your shoulder and observe the soil under the color chips on the page compare the soil color observed with the color chips in the Munsell Soil Color Chart until you determine the closest matching chip. This often requires examining a number of color pages and color chips.

When determining soil colors using the Munsell Soil Color Chart, start on the page labeled with a hue of 10YR in the upper right-hand corner. If the soil appears redder than the color chips located on the 10YR page, go toward the front of the book. If it appears yellower than the color chips on the 10YR page, go toward the back of the book. Although a range of hydric soils occur in nature, hydric soils often exhibit yellower colors compared to nonhydric soils. Also, many hydric soils contain gley colors (i.e., GLEY 1 and GLEY 2).

Color Patterns in Soil

Many soils exhibit a mixture of colors within a given soil layer. Accurately measuring the color patterns observed in soils is essential for hydric soil identification and delineation. The hydric soil data form includes spaces for recording several aspects of soil color including matrix color and abundance, redoximorphic feature color, abundance, type, and location. We briefly discuss each of these components below. See Vepraskas (2004), Vepraskas and Sprecher (1997), and Richardson and Daniels (1993) for additional insight into the formation and morphology of hydric soil colors.

1. Matrix color

   The matrix color of the soil is the dominant soil color covering the highest percentage of the soil surface area (USDA NRCS 2010). Many soil layers exhibit one matrix color as well as one or more additional colors. For example, if a soil contains two colors, one color covering 60% of the surface area and the other covering 40% of the surface area, the color accounting for 60% represents the matrix color. Once you determine the matrix color, record the hue, value, and chroma of that color on the data form under the column labeled "Matrix Color" (Figure 8.3). Additionally, record the percentage of the soil surface area occupied by the matrix color. If two dominant soil colors cover an equal surface area of the soil, record both colors as the matrix color. For example, if a soil contains three colors covering 40%, 40%, and 20% of the surface area respectively, record the matrix colors as the two colors that cover 40% of the soil surface.

2. Redoximorphic feature colors

   The term "mottles" refers to minor colors in the soil that differ from the matrix color (Soil Survey Staff 1999). Mottles created by wet conditions within the soil and the presence of anaerobic conditions are called redoximorphic features (Wakeley et al. 1996). Redoximorphic features form following the dissolution, translocation, and reprecipitation of iron and manganese within saturated and anaerobic soils (Reddy and DeLaune 2008). Redoximorphic features occur in irregular patterns along ped faces, root channels, or on the interior of peds, forming distinct patterns used for hydric soil identification (Birkeland 1999). Vepraskas (2004) provides an excellent discussion of redoximorphic feature formation, morphology, and the associated terminology. This chapter provides guidance on collecting data on redoximorphic features aimed at identifying hydric soils and completing the hydric soils data form (Figure 8.3), including the measurement of redoximorphic feature color, abundance, type, and location (Figure 8.5). We discuss each component below:

a. Redoximorphic feature colors: As described above, redoximorphic features are nonmatrix colors resulting from periods of soil wetness. These features typically include yellowish-brown, red, bluish-black, or gray colors (Birkeland 1999). Using the Munsell Soil Color Chart as described above, record the color of any redoximorphic features observed within the soil on the data form. Many hydric soils contain more than one color of redoximorphic features. For example, a hydric soil may have both yellowish-brown and gray redoximorphic features. Where this occurs, record both redoximorphic feature colors on the data form.
b. Redoximorphic feature abundance: In addition to recording redoximorphic feature color, also determine the abundance of each redoximorphic feature observed. For example, if a soil contains 60% matrix color, 25% yellowish-brown redoximorphic features, and 15% gray redoximorphic features, record the matrix color and abundance, and then record both redoximorphic features colors and their abundances utilizing the columns provided on the data form.
c. Redoximorphic feature type: The data form contains a column labeled "Type" (Figure 8.3). Redoximorphic features occur in four types including concentrations (C), depletions (D), reduced matrix (RM), and coated or masked sand grains (CS). Redoximorphic concentrations consist of areas within the soil where iron and/or manganese compounds are accumulated and concentrated, resulting in the yellow-brown, red, and bluish-black features described above (Figure 8.5; Vepraskas 2004). Redoximorphic depletions occur as whitish-gray zones within the soil where iron and/or manganese compounds have been translocated or stripped away exposing the uncoated soil grain. RM refers to the potential for saturated and reduced soils to change color following exposure to oxygen (see "Reduced Matrix" below). Masked sand grains occur when the organic material covers sandy particles with a black, greasy coating (see "Masked Sand Grains" below). Record the type of redoximorphic features observed on the data form.
d. Redoximorphic feature location: Redoximorphic features occur in two potential locations. Many redoximorphic features develop along the pore linings associated with root channels and soil ped faces (Figure 8.5). Alternatively, redoximorphic features develop within the soil matrix and are not associated with root channels or other pore linings. In many cases, hydric soils exhibit redoximorphic features in a combination of both pore linings and matrix locations; record both locations when this occurs.

Mottles resulting from things other than wetness such as mixing of soil material from adjacent horizons or weathering fragments of gravel or bedrock are not relevant to hydric soil identification. As a result, mottles caused by factors other than wetness should be noted in the remarks section of the data form and are not recorded as redoximorphic features.

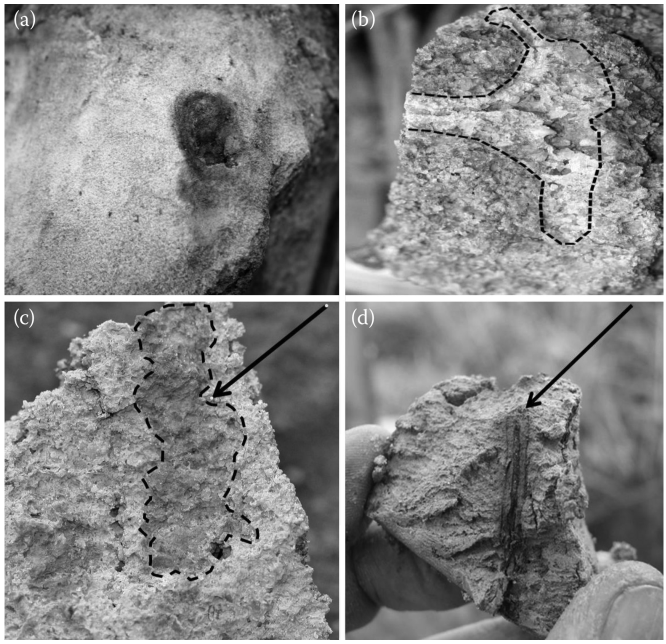


FIGURE 8.5
(a) Grayscale image of a soil exhibiting a light grey matrix color occupying 90% of the surface area and a dark redoximorphic concentration occupying 10% of the soil surface area (center-right of frame). (b) Grayscale image of a whitish-grey redoximorphic depletion (highlighted by the black dotted line) occupying 25% of the soil surface area. (c) Redoximorphic concentration located along a pore lining/root channel occupying 12% of the soil surface area. (d) Redoximorphic concentration located within the soil matrix (highlighted by the black dotted line) occupying 30% of the soil surface area.

For the purposes of the indicators, those features that are faint in contrast do not count toward the required percentage of redoximorphic features. Faint features are evident only on close examination. The contrast is faint if:

1. Delta hue = 0, then, delta value ≤2 and delta chroma ≤1, or
2. Delta hue = 1, then, delta value ≤1 and delta chroma ≤1, or
3. Delta hue = 2, then, delta value = 0 and delta chroma = 0, or
4. Any delta hue if both colors have value ≤3 and chroma ≤2.

Soil Textures for Hydric Soils Identification

Soil texture defines the physical distribution of mineral sand, silt, and clay particles within a soil sample (Singer and Munns 2002). Soil scientists classify soils into groups based on soil texture, and soil texture represents an important soil characteristic used to determine which *Field Indicators of Hydric Soils in the United States* apply in each layer of the soil. For hydric soil identification, each layer of soil is classified into one of six categories. Organic soil materials include soil textures with high organic carbon contents exhibiting various degrees of decomposition as seen in (1) peat, (2) mucky peat, or (3) muck. Mineral soil materials include soil textures dominated by (4) sandy, (5) loamy/clayey, and (6) mucky-modified mineral soil material.

1. Organic soil material—Peat, mucky peat, and muck

   The USDA NRCS (2010, p. 9) defines organic soil materials as "soil material that is saturated with water for long periods or artificially drained and, excluding live roots, has 18 percent or more organic carbon with 60 percent or more clay, or 12 percent or more organic carbon with 0 percent clay. Soils with an intermediate amount of clay have an intermediate amount of organic carbon. If the soil is never saturated for more than a few days, it contains 20 percent or more organic carbon. Organic soil material includes muck, mucky peat or peat" (Figure 8.6).

2. Mucky-modified mineral soil material

   Mucky-modified mineral soil material is defined as a mineral soil in which muck accounts for 5%–12% organic carbon with no clay, between 12% and 18% organic carbon with 60% clay, and intermediate amounts of organic carbon with intermediate amounts of clay (Figure 8.6). Mineral soils containing peat or mucky peat but lacking muck cannot qualify as mucky-modified mineral soil material (USDA NRCS 2010).

3. Mineral soil material

   Mineral soil materials contain low amounts of organic carbon, and fail to classify as peat, mucky peat, muck, or mucky-modified soil materials (Figure 8.6). Mineral soil material is predominantly composed of sand, silt, and clay (Singer and Munns 2002). Mineral soil materials classify as either sandy or loamy/clayey based on the percentage of sand, silt, and clay.

Determining Soil Texture in the Field

Soil texture determinations made in the field rely on a number of simple techniques including evaluating soil color, weight, and feel. With practice, these techniques are approachable and reliable tools used by wetland professionals, academics, and nonsoil scientists. To promote accuracy in identifying soil texture in the field, it is important to calibrate your

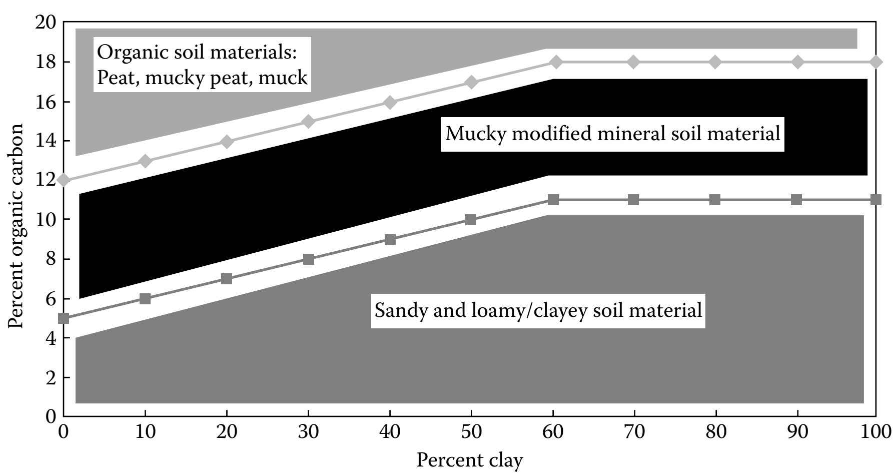


FIGURE 8.6
Soil textures are determined based on the amount of organic carbon and clay. Peat, mucky peat, and muck soils contain high amounts of organic carbon. Mineral soils contain low amounts of organic carbon, and mucky modified soil materials contain moderate amounts of organic carbon. (Adapted from USDA NRCS. 2010. Field indicators of hydric soils in the United States, Version 7.0. In L. M. Vasilas, G. W. Hurt, and C. V. Noble (Eds.) *USDA, NRCS, in cooperation with the National Technical Committee for Hydric Soils*.)

fingers using known lab-identified soil textures. Additionally, work with an experienced soil scientist familiar with the soils in your region. The following provides a step-by-step approach to making field soil texture determinations for the identification of hydric soils:

1. Determining organic versus mineral soil texture: The easiest way of identifying an organic soil is by observing the soil color, weight, and feel.
   a. Soils high in organic carbon are dark in color and typically exhibit values of 3 or less. Many organic soils that remain saturated for long periods of time also display a chroma of 2 or less.
   b. Since organic material has a lower bulk density than mineral particles, organic soil material remains notably lighter than an equivalent amount of mineral soil material.
   c. Finally, organic soil material feels greasy when rubbed between the thumb and forefinger. Conversely, mineral soil materials feel gritty if they contain sand, leave a powdery residue on your hand if it contains silt, or feel sticky if it contains clay.
2. Organic soil textures: If the soil is identified as organic soil material based on the dark color, low density, and greasy feel, the soil must be categorized as peat, mucky peat, or muck. Determining the category of organic soil material becomes important because some *Field Indicators of Hydric Soils in the United States* restrict the types of organic soil material that apply in certain situations (USDA NRCS 2010).
   a. Muck represents the most highly decomposed form of organic soil material with few or no identifiable plant structures visible in the soil; peat represents the least-decomposed form of organic soil material in which many of the plant structures (leaf veins, roots, and needles) remain readily visible. Mucky peat occurs as an intermediate stage of decomposition between muck and peat.
   b. To determine the category of organic soil material, form a golf ball-sized sample (~40 mm) of the soil material and rub it between your fingers and your thumb about 8–10 times. Break the ball of soil open and examine the abundance of visible nonliving plant fibers and roots present. The organic soil classifies as muck if less than 1/6 of the sample contains nonliving fibers (Table 8.1). The organic soil classifies as peat if more than 3/4 of the sample contains fibers. The organic soil classifies as mucky peat if between 1/6 and 3/4 of the sample contains fibers.
3. Mineral soil textures (see Chapter 1): If the soil texture is identified as mineral soil material based on the color, weight, and feel, the soil must be categorized as mucky-modified mineral soil material, sandy, or loamy/clayey.
   a. Mucky-modified mineral soil materials will be heavier than organic soil material but lighter than mineral soil material. Mucky-modified soil materials feel slightly gritty when containing sand, leave a powdery silt residue when containing silt, and feel sticky when containing clay. Also, mucky-modified mineral soil material contains highly decomposed muck resulting in a black stain on your fingers. Mucky-modified mineral textures are difficult to identify in the field, because they exhibit some components of organic soils and some components of mineral soils (Figure 8.6). Careful calibration with known texture samples proves helpful with field identifications. However, in some cases, the opinion of a qualified soil scientist or analysis at a lab is needed to confirm that the texture is in fact mucky-modified mineral.

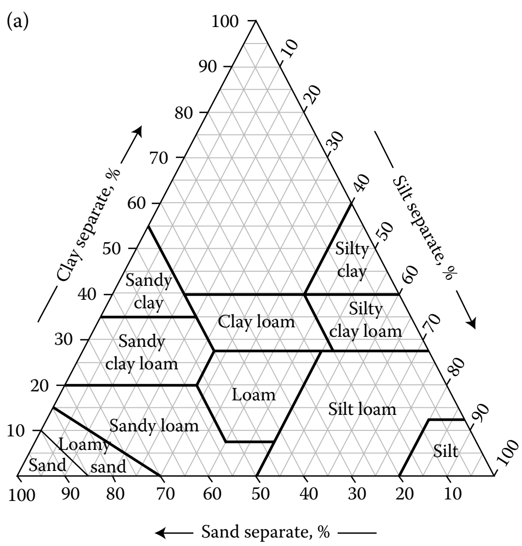
b. All mineral soil textures (including mucky-modified mineral soils) must be classified as either sandy or loamy/clayey soil material. The USDA NRCS uses additional classes of soil textures for a variety of purposes (Figure 8.7). However, for the identification of hydric soils, all mineral soils are classified as either sandy or loamy/clayey. Sandy soil textures included soils that are loamy fine sand, or coarser. Loamy/clayey textures include soils that are finer than loamy fine sand.

c. To determine the texture of mineral soils, take a golf ball-sized (~40 mm) sample of the soil and moisten. Push the soil between your fingers and thumb as described in Figure 8.8. If the soil forms a ribbon, it is loamy/clayey. The soil is sandy if it feels gritty, fails to maintain a ball when gently bounced, and fails to form a stable ribbon. At least 70% sand is required for a soil to be sandy; so, if only slight grittiness is detected and the soil feels sticky, it is most likely a loamy/clayey texture. An alternative method of determining mineral soil textures is to oversaturate the sample and mix the particles into water. Decant the muddy water containing the silt and clay, and what is left in represents the sand fraction. Estimate whether the amount of material left in the container represents a high-enough percentage of the sample to classify the soil as sandy.

TABLE 8.1

Strategy for Determining Organic Soil Texture Categories

Organic Soil Texture	Prior to 8–10 Rubs	After 8–10 Rubs
Muck	<33%	<17%
Mucky peat	33%–67%	17%–40%
Peat	>67%	>40%



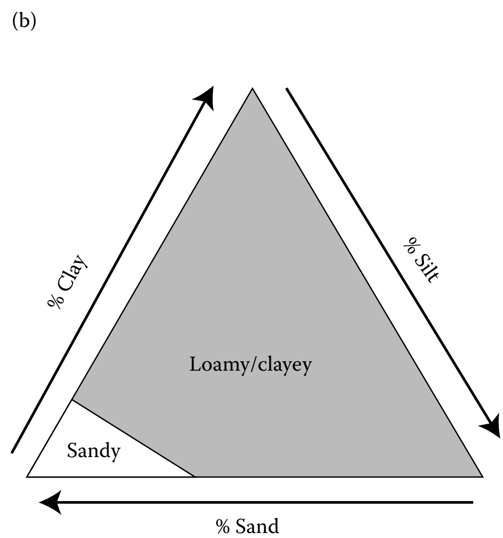


FIGURE 8.7
(a) The USDA textural triangle. (b) The "basic" soil texture triangle used for identification of hydric soils with mineral textures. (From Schoeneberger, P. J. et al. 2012. *Field Book for Describing and Sampling Soils, Version 3.0.* National Resources Conservation Service, National Soil Survey Center, Lincoln, NE.)

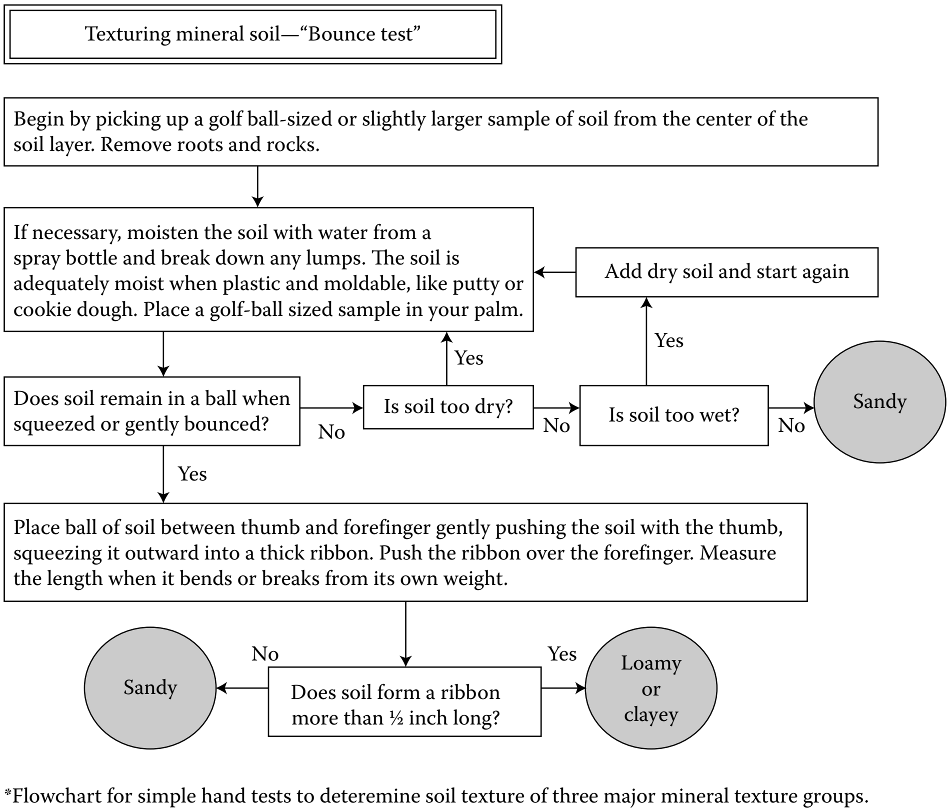


FIGURE 8.8
Basic flow chart for determining mineral soil texture for the purpose of hydric soil identification. (Modified from Thien, S. J. 1979. *J. Agron. Educ*. 8: 54–55.)

Other Important Soil Features and Concepts for Identifying Hydric Soils

The information presented above outlines the data required to complete a hydric soil data form for the purposes of hydric soil identification and delineation. Required measurements include the depth of each soil layer, matrix color and abundance, redoximorphic feature color and abundance, redoximorphic feature type and location, and soil texture (Figure 8.3). The following concepts and terminology present additional information regarding hydric soil identification. Many of the terms utilized below appear in the *Field Indicators of Hydric Soils in the United States* (USDA NRCS 2010). Major categories include morphological expressions of reduced conditions, soil matrices commonly associated with hydric soils, soil horizons of interest, and notable landscape positions and geologies. Gaining an understanding and familiarity with these terms and concepts promotes accuracy and efficiency when identifying hydric soils in a field setting.

Morphological Expressions of Reduced Conditions

As outlined above and elsewhere in the chapter, the onset of anaerobic conditions and chemical reduction leads to a number of changes in the biogeochemistry of hydric soils

(Reddy and DeLaune 2008). These changes alter the morphology of the soil and result in observable patterns and features utilized to identify and delineate hydric soils within the landscape (National Research Council 1995; Tiner 1999). The sections below address the common morphological features and terminologies used throughout the *Field Indicators of Hydric Soils in the United States* (USDA NRCS 2010).

Organic Matter Accumulation

The anaerobic and reduced conditions associated with hydric soils lead to a decrease in microbial respiration efficiency and organic matter decomposition (Schink 1988; Lee 1992). As a result, organic carbon accumulates in hydric soils (Gambrell and Patrick 1978; Mausbach and Richardson 1994). The following terms appear in the *Field Indicators of Hydric Soils in the United States* and aid in the identification of hydric soils:

1. Histosol: Histosols are defined as organic soils exhibiting 40 cm or more of organic soil material within the upper 80 cm. The 40 cm of organic soil material does not need to occur in one continuous layer. Histosols also include soils containing organic materials of any thickness if the organic soil materials are underlain by rock or fragmental materials with interstices filled with organic soil materials. See *Soil Taxonomy* (Soil Survey Staff 1999) for a complete definition (USDA NRCS 2010). Histosols are found throughout the United States. However, Histosols are common across the north-central United States, including large expanses of organic flats in northern Minnesota and large portions of southeastern and northern Alaska (Moore and Bellamy 1974; Ping et al. 1997, 2002).
2. Histic Epipedon: A Histic Epipedon is defined as a thick (20–60 cm) organic soil horizon that saturates with water during some period of the year (unless artificially drained). The organic soils materials must be underlain by a mineral soil with a chroma of 2 or less (USDA NRCS 2010). Histic Epipedons predominantly occur in northern and high-elevation regions of the United States. They are also found across many portions of the northeastern states, including marsh ecosystems as well as in coastal plain portions of the south Atlantic states (Bridgham and Richardson 1993; USDA NRCS 1997).
3. Organic bodies: The term "organic bodies" describes accumulations of highly decomposed (muck organic soil material) or mucky-modified mineral soil material occurring at the tips of fine roots. To identify organic bodies, remove a large clump of roots from the soil and gently shake to remove loose soil material. If clumps of soil material remain on the tips of fine roots, rub them between the thumb and forefinger to determine if they are muck (Table 8.1). If the organic material present is classified as muck or mucky-modified mineral soil material, the soil would qualify as organic bodies. Organic bodies occur in the southern Gulf of Mexico and the Florida panhandle as well as other areas along the Atlantic coast. These features are often associated with wet pine flat landscapes (Florida DEP 2011).
4. Masked sand grains: Sand grains may become coated or masked with organic material under anaerobic and reduced soil conditions associated with extended periods of saturation (USDA NRCS 2010). Organic matter covers the sand particle in black organic material, hiding the original soil color. Other potential masking agents include silicate clays, iron, aluminum, or some combination of these;

however, organic matter masking remains most common in hydric soils (Lindbo et al. 2000). Typically, if at least 70% of sand grains in a sandy soil appear masked with organic matter, that soil is considered to undergo extended periods of saturation and reduced conditions. When examining coated or masked sand grains, utilize a 10× magnitude hand lens to determine if the 70% threshold is met or exceeded. When observed with the naked eye, the percentage of masked grains appears close to 100% masked.

Reduction and Translocation of Iron, Manganese, and Sulfur

The anaerobic and reduced conditions associated with hydric soils result in the chemical reduction and translocation of iron, manganese, and sulfur (Reddy and DeLaune 2008). These elements may reprecipitate following the onset of aerobic conditions associated with decreasing water tables (Megonigal et al. 1993, 1996; Vepraskas 2004). As a result, iron and manganese often develop a characteristic pattern of redoximorphic features associated with hydric soils (Vepraskas and Sprecher 1997). Sulfur reduction results in an olfactory indication of chemical reduction (Castro and Dierberg 1987). Together, the reactions of these elements and compounds, and associated terminology, aid in the identification of hydric soils.

1. Redoximorphic concentrations as iron, manganese, or iron and manganese: As described above, redoximorphic concentrations occur where bodies of iron and/or manganese accumulate through the process of dissolution, translocation, and reprecipitation (Fanning and Fanning 1989). Concentrations of iron are typically red, orange, brown, or yellow in color depending on the form of iron, while concenrations of manganese are typically black in color (Birkeland 1999). If the iron and manganese concentrate together, the color will be blackish red, purple, or black in color. Some *Field Indicators of Hydric Soils in the United States* do not specify the type of redoximorphic concentration; however, others require the presence of manganese, iron, or a combination of iron and manganese features to be present (USDA NRCS 2010). If you are unsure if the redox concentrations observed within a soil contain manganese, a solution of 2% hydrogen peroxide can be placed onto the soil. If the application of hydrogen peroxide causes the soil to effervesce, the redoximorphic concentration contains manganese (Needelman et al. 2007).
2. The use of α-α dipyridyl dye: The compound α-α dipyridyl is a colorless liquid dye that produces a pink or red color when placed on a saturated and chemically reduced soil containing ferrous iron (NTCHS 2009). The dye provides a positive indicator of reducing conditions in a soil if a reaction is observed. However, due to several limitations associated with the dye (e.g., soils must be saturated prior to application, uncertain shelf life of the dye), a lack of a reaction does not provide a negative indicator of reducing conditions or the presence of a hydric soil (Vepraskas 2004).
3. Hydrogen sulfide odor: In soils subject to extended periods of saturation and anaerobiosis, a variety of sulfur compounds become reduced and hydrogen sulfide gas is produced (Hedin et al. 1989). The odor of hydrogen sulfide gas provides another indicator of anaerobic and chemically reduced conditions in soils.

Hydrogen sulfide exhibits a strong odor, most often described as the smell of rotten eggs (USDA NRCS 2010). Hydrogen sulfide odors often occur in coastal and tidal areas where regular inundation occurs throughout the tidal cycle and ample sulfur compounds exist (Koch et al. 1990). The occurrence of hydrogen sulfide odor in the presence of anaerobic conditions in the soil results in an obvious odor that can often be detected prior to excavation or shortly after excavation of the soil pit. If an odor is observed only when the soil is held closely to the nose, it is not hydrogen sulfide odor. Also, hydrogen sulfide odor can only exist in the presence of saturated soils and anaerobic conditions. Therefore, the hydrogen sulfide odor will not occur in unsaturated soils.

Common Hydric Soil Matrices

Depleted Matrix

The term "depleted matrix" refers to the volume of a soil horizon or subhorizon in which the processes of reduction and translocation resulted in the removal or transformation of iron and manganese. These conditions create soil colors exhibiting low chroma and high value (Figure 8.9). A depleted matrix exhibits similar characteristics to redoximorphic depletions in which the dominant matrix color is represented by large

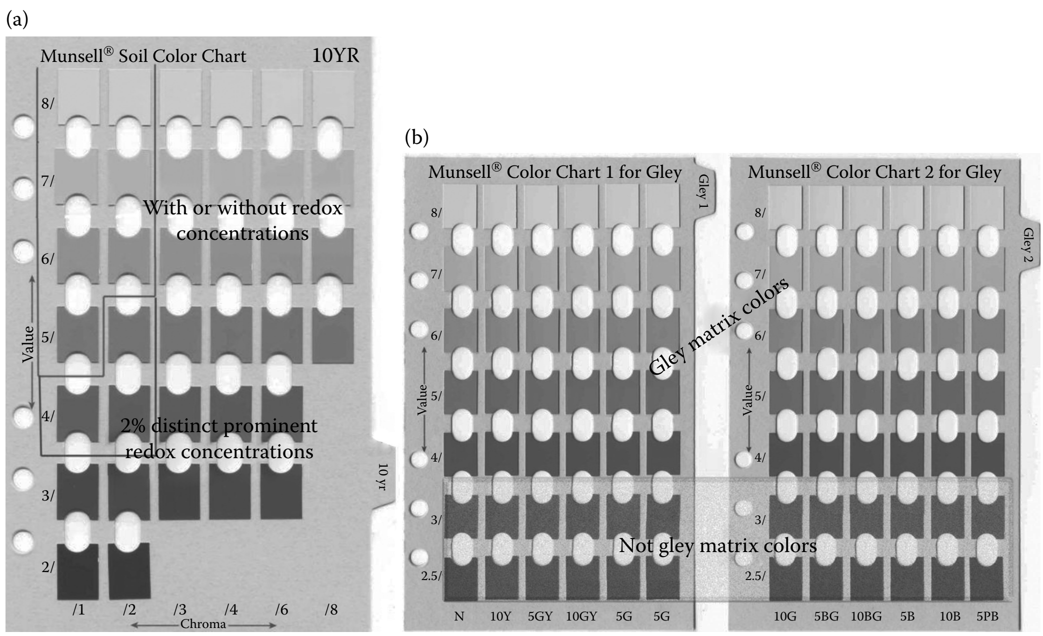


FIGURE 8.9
(a, left) Grayscale image depicting the definition of a depleted matrix, which includes color chips located with the boxes overlying the Munsell® Soil Color Chart page. Note that value/chroma combinations of 4/1, 4/2, and 5/1 require at least 2% redoximorphic concentrations, remaining color combinations require no redoximorphic features. The 10YR page is used as an example, however the same pattern applies to all other Munsell® Soil Color Chart except the gley pages. (b, middle and right) Grayscale image depicting the definition of a gleyed matrix. Note that Munsell® Soil Color Chart chips on the gley pages must have a value of 4 or more to qualify as a gley matrix.

redoximorphic depletions. In some cases, the depleted matrix changes color upon exposure to air (see "Reduced Matrix" below); this phenomenon is included in the concept of depleted matrix. Figure 8.9 and the following combinations of value and chroma identify a depleted matrix:

1. Matrix value of 5 or more and chroma of 1 or less with or without redox concentrations occurring as soft masses and/or pore linings; or
2. Matrix value of 6 or more and chroma of 2 or less with or without redox concentrations occurring as soft masses and/or pore linings; or
3. Matrix value of 4 or 5 and chroma of 2, and 2% or more distinct or prominent redox concentrations occurring as soft masses and/or pore linings; or
4. Matrix value of 4 and chroma of 1, and 2% or more distinct or prominent redox concentrations occurring as soft masses and/or pore linings (USDA NRCS 2010)

Many A, E, and calcic horizons have low chromas and high values, and may therefore be mistaken for a depleted matrix; however, they are excluded from the concept of depleted matrix unless the soil has common or many distinct or prominent redox concentrations occurring as soft masses or pore linings (USDA NRCS 2010).

Gleyed Matrix

Gley soils occur in areas exposed to saturation for periods of significant duration, allowing for the reduction and translocation of large amounts of iron and manganese (USDA NRCS 2010). Figure 8.9 and the color combinations listed below describe the gleyed matrix:

1. 10Y, 5GY, 10GY, 10G, 5BG, 10BG, 5B, 10B, or 5PB with a value of 4 or more and chroma of 1; or
2. 5G with a value of 4 or more and chroma of 1 or 2; or
3. N with a value of 4 or more

In some cases, the gleyed matrix may change color upon exposure to (see "Reduced Matrix" below). When a reduced matrix occurs in a soil initially displaying a gley matrix, the soil is included in the concept of a gleyed matrix (USDA NRCS 2010). The concept of a gleyed matrix does not include all glauconitic soils, which display grayish-green colors potentially misidentified as a gleyed matrix. When encountering glauconitic soil materials, examine the soils for redoximorphic features and, if needed, consult with an experienced soil scientist (see "Glauconite" below).

Reduced Matrix

If the soil appears saturated at the time of excavation, collect soil color measurements immediately upon excavation. Carefully examine the soil to determine if a color change occurs after several minutes of exposure to air. Record the change in color if observed. The change in color indicates the presence of reduced iron in the soil solution at the time of excavation and is known as a "reduced matrix" (Vepraskas 2004). The reduced iron subsequently oxidizes upon exposure to oxygen, resulting in a change in color. This only occurs in saturated and chemically reduced soils. See USACE (2012) for a detailed description of the reduced matrix and additional instructions for recording observed color changes on the hydric soil data form and the amount of time it takes for the color to change.

Other Soil Horizon Characteristics of Interest

A, E, and Calcic Horizons

Soil scientists generally collect additional soil information beyond what is required for hydric soil identification, including the assignment of soil horizon designations. These soil horizons designations describe processes that are occurring in that soil horizon. It is important to be able to recognize A, E, and calcic horizons because these soil horizons can be misidentified as a depleted matrix (USDA NRCS 1996). For example, A horizons typically occur in the uppermost mineral horizon(s) in the soil profile. A horizons are dark in color due to organic matter accumulation. E horizons typically occur immediately below an A horizon and exhibit pale colors due to a loss of organic matter, iron, clay, and other minerals from weathering. These pale soil colors can potentially be misidentified as a depleted matrix. Calcic horizons are layers in the soil where significant amounts of carbonates accumulate, resulting in pale gray or white soil colors. These soil horizons are excluded from the concept of depleted matrix unless the soil contains a minimum of 2% redox concentrations.

Stratified Layers

While all soils develop in layers, the term "stratified layers" used for hydric soil identification refers to soils exhibiting pedogenic discontinuity. This means that soil layers containing high concentrations of organic matter have been buried below newly deposited sediments. Stratified layers typically occur in active floodplains where a surface layer of organic soil material, mucky-modified mineral material, or dark-colored mineral soil materials has been buried under alluvial sediments.

Spodic Horizon

Some *Field Indicators of Hydric Soils in the United States* contain additional requirements if the soil contains a spodic horizon. A spodic horizon refers to a mineral soil horizon characterized by the accumulation of amorphous materials consisting of aluminum, organic carbon, and potentially iron (USDA NRCS 2010). Spodic horizons display dark reddish or coffee brown or black colors. Spodic horizons typically occur below an E horizon. See *Soil Taxonomy* (Soil Survey Staff 1999) for a more complete definition of spodic horizons.

Notable Landscapes and Geologies

Some *Field Indicators of Hydric Soils in the United States* limit application or exclude specific landscapes and/or geologies. An understanding of the following landscapes and geologies is important for hydric soil identification.

Closed Depression

Several of the *Field Indicators of Hydric Soils in the United States* require that the sampling location takes place within a closed depression or a closed depression subject to ponding.

A closed depression is defined as a low-lying area that is surrounded by a higher ground and has no natural outlet for surface drainage (USDA NRCS 2010). If the depression must also be subject to ponding, evidence that the water table rises above the surface must also be present. While a depression may not have a natural outlet in instances of very high precipitation events, flooding water may exit the depression through nonpoint overland flow. Closed depressions occur in many landscape settings. Closed depressions subject to ponding are frequently located in backwater depressions on floodplains but also occupy depressions in flat landscapes such as prairie potholes, Delmarva Bays, Grady ponds, and Carolina bays (Schalles and Shure 1989; Sharitz 2003).

Floodplains

A floodplain is a "nearly level plain that borders a stream and is subject to inundation under flood-stage conditions unless protected artificially. It is usually a constructional landform built of sediment deposited during overflow and lateral migration of the streams." A floodplain is not a stream terrace that is "one, or a series of flat-topped landforms in a stream valley that flank and are parallel to the stream channel, originally formed by a previous stream level, and representing remnants of an abandoned flood plain, stream bed, or valley floor produced during a past state of fluvial erosion or deposition (i.e., currently very rarely or never flooded; inactive cut and fill and/or scour and fill processes). Erosional surfaces cut into bedrock and thinly mantled with stream deposits (alluvium) are called 'strath terraces.' Remnants of constructional valley floors thickly mantled with alluvium are called alluvial terraces" (USDA NRCS 2013, p. 5, 10).

Red Parent Material

Red parent material is defined as parent materials with a natural inherent reddish color attributable to the presence of iron oxides, typically hematite (Elless and Rabenhorst 1994; Elless et al. 1996), occurring as coatings on and occluded within mineral grains. Soils that formed in red parent material exhibit conditions that retard the development and extent of the redoximorphic features that normally occur under prolonged aquic conditions. They typically display a Color Change Propensity Index (CCPI) of <30 (Rabenhorst and Parikh 2000). Most commonly, the material consists of dark red, consolidated Mesozoic or Paleozoic sedimentary rocks, such as shale, siltstone, and sandstone, or alluvial materials derived from such rocks. Assistance from a local soil scientist may help to determine where the red parent material occurs (USDA NRCS 2010).

Glauconite

The term "glauconite" refers to a mineral aggregate that contains a micaceous mineral resulting in a characteristic green color, for example, glauconitic shale or clay (USDA NRCS 2010). The concept of a gleyed matrix does not include all glauconitic soils, which can be potentially misidentified as a gleyed matrix. When encountering glauconitic soil materials, examine the soils for redoximorphic features and, if needed, consult with an experienced soil scientist. When identifying redoximorphic features in soils containing glauconite, be cautious of orange or yellow mottles that result from the oxidation of sulfur that is not related to anaerobic conditions.

Marl

Marl is an earthy, unconsolidated deposit chiefly consisting of calcium carbonate mixed with clay in approximately equal proportions. It is primarily formed under freshwater lacustrine conditions (USDA NRCS 2010). See *Soil Taxonomy* (Soil Survey Staff 1999) for a more complete definition.

References

Bigham, J. M. and E. J. Ciolkosz. 1993. *Soil Color.* Soil Science Society of America Publication Number 31, Soil Science Society of America, Madison, WI.

Birkeland, P. W. 1999. *Soils and Geomorphology*. Oxford University Press, New York.

Bridgham, S. D. and C. J. Richardson. 1993. Hydrology and nutrient gradients in North Carolina peatlands. *Wetlands* 13: 207–218.

Castro, M. S. and F. E. Dierberg. 1987. Biogenic hydrogen sulfide emissions from selected Florida wetlands. *Water Air Soil Pollut.* 33: 1–13.

Cox, C. B. and P. D. Moore. 1993. *Biogeography: An Ecological and Evolutionary Approach*, 5th ed. Blackwell Scientific, London.

Daniels, R. B., E. E. Gamble, and J. G. Cady. 1971. The relation between soil morphology and genesis. *Adv. Agron.* 23: 51–88.

Elless, M. P. and M. C. Rabenhorst. 1994. Hematite in the shales of the triassic Culpeper basin of Maryland. *Soil Sci.* 158: 150–154.

Elless, M. P., M. C. Rabenhorst, and B. R. James. 1996. Redoximorphic features in soils of the triassic Culpeper Basin. *Soil Sci.* 161: 58–69.

Environmental Laboratory. 1987. *U.S. Army Corps of Engineers Wetland Delineation Manual*. Technical Report Y-87-1. Army Engineer Waterways Experiment Station, Vicksburg, MS.

Fanning, D. E. and M. C. B. Fanning. 1989. *Soil: Morphology, Genesis, and Classification*. John Wiley and Sons, New York.

Faulkner, S. F. 2004. Soils and sediments. In S. L. Spray, and K. L. McGlothlin (Eds.) *Wetlands*. Rowman and Littlefield Publishers, Inc., Washington, DC.

Federal Register. July 1994. Changes in hydric soils of the United States. Volume 59, Number 133 (Wednesday, July 13, 1994).

Florida Department of Environmental Protections (DEP). 2011. *Soil and Water Relationships of Florida's Ecological Communities*. Florida Department of Environmental Protection, Soil Survey Staff, Tallahassee, Florida.

Gambrell, R. P. and W. H. Patrick. 1978. Chemical and microbiological properties of anaerobic soils and sediments. In D. D. Hook, and M. M. Crawford (Eds.) *Plant Life in Anaerobic Environments*. CRC Press, Ann Arbor, MI.

Gretag/Macbeth. 2000. *Munsell® color*. New Windsor, NY.

Hedin, R. S., R. Hammack, and D. Hyman. 1989. Potential importance of sulfate reduction processes in wetland constructed to treat mine drainage. In D. A. Hammer (Ed.) *Constructed Wetlands for Wastewater Treatment*. CRC Press, Boca Raton, Florida.

Koch, M. S., I. A. Mendelssohn, and K. L. McKee. 1990. Mechanism for the hydrogen sulfide induced growth limitation in wetland macrophytes. *Limnol. Oceanogr.* 35: 399–408.

Lee, C. 1992. Controls on organic carbon preservation: The use of stratified water bodies to compare intrinsic rates of decomposition in oxic and anoxic systems. *Geochim. Cosmochim. Acta* 56: 3323–3336.

Lindbo, D. L., M. J. Vepraskas, and F. E. Rhoton. 2000. A field method for determining percent coated sand grains. *Soil Sci. Soc. Am. J.* 65: 949–953.

Mausbach, M. J. and J. L. Richardson. 1994. Biogeochemical processes in hydric soil formation. *Current Topics in Wetland Biogeochemistry*. Louisiana State University, Wetland Biogeochemistry Institute, Baton Rouge, Louisiana.

Megonigal, J. P., S. P. Faulkner, and W. H. Patrick. 1996. The microbial activity season in southeastern hydric soils. *Soil Sci. Soc. Am. J.* 60: 1263–1266.

Megonigal, J. P., W. H. Patrick, and S. P. Faulkner. 1993. Wetland identification in seasonally flooded forest soils: Soil morphology and redox dynamics. *Soil Sci. Soc. Am. J.* 32: 634–638.

Mitsch, J. M. and J. S. Gosselink. 2007. *Wetlands*, 4th ed. John Wiley and Sons, Hoboken, New Jersey.

Moore, P. D. and D. J. Bellamy. 1974. *Peatlands*. Springer-Verlag, New York.

National Research Council. 1995. *Wetlands: Characteristics and Boundaries.* National Academy Press, Washington, DC.

National Technical Committee for Hydric Soils (NTCHS). 2009. *Use of Alpha-Alpha Dipyridyl. Hydric Soils Technical Note 8.* National Technical Committee for Hydric Soils, United States Department of Agriculture, Natural Resources Conservation Service, Washington, DC.

National Technical Committee for Hydric Soils (NTCHS). 2012. *Hydric Soil List Criteria*. United States Department of Agriculture, Natural Resources Conservation Service, Washington, DC.

Needelman, B. A. and D. E. Ruppert Vaughan. 2007. The role of ditch soil formation and redox biogeochemistry in mitigating nutrient and pollutant losses from agriculture. *J. Soil Water Conserv.* 62: 207–215.

Ping, C. L., G. L. Michaelson, and J. M. Kimble. 1997. Carbon storage along a latitudinal transect in Alaska. *Nutrient Cycl. Agroecosyst.* 49: 235–242.

Ping, C. L., G. J. Michaelson, J. M. Kimble, and R. Everett. 2002. Organic carbon stores in Alaska soils. In R. Lal, J. M. Kimble, and R. F. Follett (Eds.) *Agriculture Practices and Policies for Carbon Sequestration in Soil*. CRC Press, Boca Raton, Florida.

Post, D. F., S. J. Levine, R. B. Bryant, M. D. Mays, A. K. Batchily, R. Escadafal, and A. R. Huete. 1993. Correlations between field and laboratory measurements of soil color. In J. M. Bigham, and E. J. Ciolkosz (Eds.) *Soil Color.* Soil Science Society of America Publication Number 31, Soil Science Society of America, Madison, WI.

Rabenhorst, M. C. and S. Parikh. 2000. Propensity of soils to develop redoximorphic color changes. *Soil Sci. Soc. of Am. J.* 64: 1904–1910.

Reddy, K. R. and R. D. DeLaune. 2008. *Biogeochemistry of Wetlands*. CRC Press, Boca Raton, Florida.

Richardson, J. L. and R. B. Daniels. 1993. Stratigraphic and hydraulic influences on soil color development. In *Soil Color.* Soil Science Society of America Publication Number 31, Soil Science Society of America, Madison, WI.

Sahrawat, K. L. 2003. Organic matter accumulation in submerged soils. *Adv. Agron.* 81: 161–209.

Schalles, J. F. and D. J. Shure. 1989. Hydrology, community structure, and productivity patterns of a dystrophic Carolina Bay wetland. *Ecol. Monogr.* 59: 365–385.

Schink, B. 1988. Principles and limits of anaerobic degradation: Environmental and technological aspects. In A. J. B. Zehnder (Ed.) *Biology of Anaerobic Microorganisms*. Wiley, New York.

Schoeneberger, P. J., D. D. Wysocki, E. C. Benton, and Soil Survey Staff. 2012. *Field Book for Describing and Sampling Soils, Version 3.0*. National Resources Conservation Service, National Soil Survey Center, Lincoln, NE.

Sharitz, R. R. 2003. Carolina Bay Wetlands: Unique habitats of the southeastern United States. *Wetlands* 23: 550–562.

Singer, M. J. and D. N. Munns. 2002. *Soils: An Introduction*, 5th ed. Prentice-Hall, Upper Saddle River, New Jersey.

Soil Survey Staff. 1951. *Soil Survey Manual. U.S. Department of Agriculture Handbook 18*. National Resources Conservation Service, National Soil Survey Center, Lincoln, NE.

Soil Survey Staff. 1999. *Soil Taxonomy. Agriculture Handbook 436,* 2nd ed. U.S. Department of Agriculture, Natural Resource Conservation Service, Lincoln, NE.

Thien, S. J. 1979. A flow diagram for teaching texture-by-feel analysis. *J. Agron. Educ.* 8: 54–55.

Tiner, R. W. 1999. *Wetland Indicators: A Guide to Wetland Identification, Delineation, Classification, and Mapping*. CRC Press, Boca Raton, Florida.

USACE. 2012. Hydric soil indicators. In J. S. Wakeley, R. W. Lichvar, C. V. Noble, and J. F. Berkowitz (Eds.) *Regional Supplement to the Corps of Engineers Wetland Delineation Manual: Northcentral and Northeast Region (Version 2.0)*. ERDC/EL TR-08-28. U.S. Army Engineer Research and Development Center, Vicksburg, MS.

USDA. 1985. *Hydric Soils of the United States*. Soil Conservation Service, Washington, DC.

USDA NRCS. 1993. *Soil Survey Manual. U.S. Department of Agriculture Handbook 18*. Soil Survey Division Staff, Washington, DC.

USDA NRCS. 1996. *Soil Survey Laboratory Methods Manual*. Soil Survey Division Staff, Soil Survey Investigations Report Number 42, Version 3.0, Washington, DC.

USDA NRCS. 1997. *Hydrology Tools for Wetland Determination*. Chapter 19, *Engineering Field Handbook*. U.S. Department of Agriculture, Washington, DC.

USDA NRCS. 2010. Field Indicators of Hydric Soils in the United States, Version 7.0, 2013 Errata. In L. M. Vasilas, G. W. Hurt, and C. V. Noble (Eds.) *USDA, NRCS, in cooperation with the National Technical Committee for Hydric Soils*. Washington, DC.

Vasilas, L. M. and B. L. Vasilas. 2004. *A Guide to Hydric Soils in the Mid-Atlantic Region, Version 1.0*. U.S. Department of Agriculture Natural Resources Conservation Service, Morganitown, WV.

Vepraskas, M. J. 2004. *Redoximorphic Features for Identifying Aquic Conditions*. North Carolina Agricultural Research Service Technical Bulletin 301, North Carolina State University, Raleigh, North Carolina.

Vepraskas, M. J. and S. W. Sprecher. 1997. Overview of aquic conditions and hydric soils. In M. J. Vepraskas, and S. W. Sprecher (Eds.) *Aquic Conditions and Hydric Soils*. Soil Science Society of America, Madison, WI.

Wakeley, J. S., S. W. Sprecher, and W. C. Lynn. 1996. *Preliminary Investigation of Hydric Soil Hydrology and Morphology in the United States*. Wetland Research Center Technical Report WRP-DE-13. Waterways Experiment Station, Vicksburg, MS.

9

Delineating Hydric Soils

G. Wade Hurt and Christopher V. Noble

CONTENTS

Introduction

For centuries wetlands were regarded as little more than habitat for mosquitoes, snakes, and other pests. Today, in addition to recognizing wetlands as habitats for a variety of wildlife species (including mosquitoes and snakes), we are aware that wetlands are the nursery grounds for our fisheries, filter pollutants, reduce flooding, protect against erosion, provide timber products, recharge groundwater reserves, and furnish society with educational, scientific, recreational, and aesthetic benefits. Local, state, and federal governments have enacted laws that regulate the use of wetlands to preserve these public benefits.

To be regulated, wetlands must first be identified and delineated. Most regulated wetlands under federal jurisdiction must have three essential components: (1) hydrophytic vegetation, (2) hydric soils, and (3) wetland hydrology (Cowardin et al. 1979; Environmental Laboratory 1987; Hammer 1992; Tiner and Burke 1995). Technical criteria for each of these characteristics must be met before an area can be identified as a wetland (Environmental Laboratory 1987; U.S. Department of Agriculture 1994). When anaerobic conditions prevail in wetland soils for long enough periods during the growing season, a predominance of hydrophytic vegetation is favored. Undrained hydric soils with natural vegetation should support a dominant population of ecologically facultative wetland and obligate wetland plant species; conversely, drained hydric soils without natural vegetation have the ability to support a dominant population of ecologically facultative wetland and obligate wetland plant species once hydrologic modifications are removed or are not maintained.

This chapter presents approaches and methods for identifying and delineating hydric soils for purposes of implementing Section 404 of the Clean Water Act (CWA) and the Food Security Act (FSA) of 1985 as amended by the Food, Agriculture, Conservation, and Trade Act of 1990, the Federal Agriculture Improvement and Reform Act of 1996, the Farm Security and Rural Investment Act of 2002, and the Food Conservation and Energy Act of 2008. It is designed to assist readers in making wetland determinations and delineations using hydric soils as the primary factor. Separate sections are devoted to preliminary off-site investigations and detailed examination and delineation procedures, with a special section on problem hydric soil delineations. This chapter also includes our observations and recommendations on delineating hydric soils that have been developed over 40 years of studying wetlands.

Wetland Components

Hydrology

It is recognized that the influence of water is the key parameter in the presence or absence of wetlands. Unfortunately, annual and seasonal variations in hydrology make the direct measurement of this parameter for delineating wetlands in the field very difficult, time consuming, and costly. In addition, requirements for recognition of wetland hydrology vary among regulating agencies. The U.S. Army Corps of Engineers (USACE) (2005), Environmental Protection Agency (EPA) (2005), and Natural Resources Conservation Service (NRCS) (2012) require inundation and/or saturation of the soil surface for 14 or more consecutive days of the growing season. Faulkner et al. (1991) found that more than 14 days to as many as 28 days of surface saturation annually may be required to induce sufficient anaerobic conditions for developing hydric soil morphology.

Hydrological records of several years may be required to accurately assess the hydrology of a site. Skaggs et al. (1994) documented a site that required 48 years of data to determine that wetland hydrology was present in 24 of those years and therefore met wetland requirements. However, wetland hydrology was not present for the other 24 years, and several consecutive years lacked wetland hydrology. These types of data (years to decades) are desirable but rarely available for borderline sites and most delineation edges. Determination of the hydrologic status of a site must be made based on indicators of wetland hydrology and, sometimes, short-term saturation records. The reliability of short-term saturation monitoring to determine whether wetland hydrology exists for a given site is suspect (Skaggs et al. 1991). However, Sumner et al. (2009) showed that it is possible to use short-term (e.g., 1 year) data for hydrology assessment if rainfall is properly evaluated.

Vegetation

Presence or absence of wetland vegetation is based on a dominance of plants from a list of plants that have been identified as likely to be found in wetlands (Lichvar and Kartesz 2009) that grow in areas with insufficient concentration of oxygen for root respiration. Lichvar and Minkin (2008) recognized four types of indicator plants that occur in wetlands: (1) obligate wetland plants (OBL) that almost always occur in wetlands and rarely in uplands, (2) facultative wetland plants (FACW) that usually occur in wetlands but occasionally occur in uplands, (3) facultative plants (FAC) that commonly occur in wetlands and uplands, and (4) facultative upland plants (FACU) that occasionally occur in wetlands but usually occur in uplands.

Hydric Soils

Soils provide a reliable method of delineating wetlands, especially in areas with unreliable or unavailable hydrology data in areas of transitional vegetation, or in areas where use of the plant list does not provide delineation assistance (Florida Soil Survey Staff 1992; Hurt and Brown 1995; Segal et al. 1995). According to *Field Indicators of Hydric Soils in the United States* (Field Indicators) (USDA NRCS 2010): "Nearly all hydric soils exhibit characteristic morphologies that result from repeated periods of saturation and/or inundation for more than a few days. Soil saturation or inundation activates microbiological activity that results in a depletion of oxygen. The resulting anaerobiosis promotes biogeochemical processes such as the accumulation of organic matter and the reduction, translocation, and/or accumulation of iron and other reducible elements." These processes are responsible for the formation of characteristic soil morphologies that persist during both wet and dry periods, making them particularly useful for identifying hydric and other wet soils (Mausbach and Richardson 1994; Vepraskas 1994).

The Hydric Soil Definition (*Federal Register,* July 13, 1994) is: "A hydric soil is a soil that formed under conditions of saturation, flooding, or ponding long enough during the growing season to develop anaerobic conditions in the upper part." Criteria for hydric soils were updated in 2012 (*Federal Register*, February 29, 2012). Relationships and limitations of the hydric soil definition, criteria, and field indicators must be thoroughly understood to facilitate accurate identification and delineation of hydric soils in the field. All hydric soils must satisfy the requirements of the hydric soil definition. This means the soils must be saturated or inundated during the growing season, and the soil must experience anaerobic conditions. The National Technical Committee for Hydric Soils (NTCHS) has approved that soil inundation (ponding or flooding) can be used to document the presence of a hydric soil if data proving that inundation for 7 or more consecutive days during the growing season is available and that anaerobic conditions occur as required by the hydric soil technical standard (HSTS) exist. Technical requirements of the HSTS are available at http://soils.usda.gov/use/hydric/ as technical note 11.

Presence of one (or more) field indicators (Table 9.1) as previously explained (Chapter 8) is evidence that the definition has been met because field indicators form in soils that are saturated or inundated and become anaerobic within 30 cm (12 in.) of the surface. The "growing season" is considered to be that part of the year during which above-ground vascular plants grow and develop or the soil temperature and moisture conditions permit microbial activities (Chapter 5). In soils that lack one of the field indicators anaerobic conditions and saturated conditions as defined by the HSTS must exist.

Preliminary Off-Site Investigations

Prior to any onsite identification or delineation of hydric soils, all available offsite information should be evaluated. Offsite information available for most nonfederal lands in the United States and Puerto Rico includes the published soil surveys of the National Cooperative Soil Survey, National Wetlands Inventory (NWI) Maps produced by the U.S. Fish and Wildlife Service (USFWS), the topographic quadrangle series of maps produced by the U.S. Geological Survey (USGS), and maps of areas subject to flooding produced by the Federal Emergency Management Agency (FEMA). Reviewing these sources before attempting to identify or delineate hydric soils can significantly reduce time spent in the field. It will also facilitate most onsite identification and delineation procedures.

Published Soil Surveys

The published soil survey is an excellent place to start offsite investigation before making onsite wetland determinations. Soil surveys have been completed for more than 90% of the private and nonfederal lands in the continental United States. Most of these have been published at a scale of 1:12,000–1:24,000 at the local or county level. Many soil surveys can be found online from the Web Soil Survey. This web site provides the user ability to create a variety of interpretive maps of potential hydric soil map units. However, the same limitations apply to these interpretative maps as described for soil surveys. First, the scale limitation must be considered. Most soil surveys do not show soil bodies that are less than about 1.2 hectares in size. Finally, most soil surveys were produced prior to development of the hydric soil concept.

Hydric Soil Lists

Hydric soil lists are also available at the local or county level. These lists contain soil survey map units that have a strong probability of being hydric. They were developed by comparing the estimated soil properties found in a published soil survey with specific soil criteria. Hydric soil lists have the same limitations as the soil survey and must be used with caution. The presence of a soil on a hydric soil list does not mean that it is in fact hydric; this is only an interpretive rating and must be verified in the field. If any portion of

TABLE 9.1

Field Indicators of Hydric Soils of the United States[a]

A. Field indicators for all soils regardless of texture:

A1 (Histosol or Histel)—Classifies as a Histosol, except Folist or as a Histel, except Folistel.

A2 (Histic Epipedon)—A histic epipedon underlain by mineral soil material with chroma of 2 or less.

A3 (Black Histic)—A layer of peat, mucky peat, or muck 20 cm (8 in.) or more thick that starts within the upper 15 cm (6 in.) of the soil surface; has hue of 10YR or yellower, value of 3 or less, and chroma 1 or less; and is underlain by mineral soil material with chroma of 2 or less.

A4 (Hydrogen Sulfide)—A hydrogen sulfide odor within 30 cm (12 in.) of the soil surface.

A5 (Stratified Layers)—Several stratified layers starting within the upper 15 cm (6 in.) of the soil surface. At least one of the layers has value of 3 or less with chroma 1 or less, or it is muck, mucky peat, peat or mucky modified mineral texture. The remaining layers have chroma of 2 or less. For any sandy material that constitutes the layer with value 3 or less and chroma 1 or less, at least 70% of the visible soil particles must be masked with organic material, viewed through a 10x or 15x hand lens. Observed without a hand lens, the particles appear to be close to 100% masked.

A6 (Organic Bodies)—Presence of 2% or more organic bodies of muck or a mucky modified mineral texture starting within 15 cm (6 in.) of the soil surface.

A7 (5 cm Mucky Mineral)—A layer of mucky modified mineral 5 cm (2 in.) or more thick, starting within 15 cm (6 in.) of the soil surface.

A8 (Muck Presence)—A layer of muck with value of 3 or less and chroma of 1 or less, starting within 15 cm (6 in.) of the soil surface.

A9 (1 cm Muck)—A layer of muck 1 cm (0.5 in.) or more thick with value of 3 or less and chroma of 1 or less and starting within 15 cm (6 in.) of the soil surface.

A10 (2 cm Muck)—A layer of muck 2 cm (0.75 in.) or more thick with value of 3 or less and chroma of 1 or less starting within 15 cm (6 in.) of the soil surface.

A11 (Depleted Below Dark Surface)—A layer with a depleted or gleyed matrix that has 60% or more chroma of 2 or less, starting within 30 cm (12 in.) of the soil surface, and having a minimum thickness of either:

a. 15 cm (6 in.), or
b. 5 cm (2 in.) if the 5 cm consists of fragmental soil material (see Glossary).

Loamy/clayey layer(s) above the depleted or gleyed matrix must have value of 3 or less and chroma of 2 or less. Any sandy material above the depleted or gleyed matrix must have value of 3 or less and chroma of 1 or less, and viewed through a 10x or 15x hand lens, at least 70% of the visible soil particles must be masked with organic material. Observed without a hand lens, the particles appear to be close to 100% masked.

A12 (Thick Dark Surface)—A layer at least 15 cm (6 in.) thick with a depleted or gleyed matrix that has 60% or more chroma of 2 or less starting below 30 cm (12 in.) of the surface. The layer(s) above the depleted or gleyed matrix must have value of 2.5 or less and chroma of 1 or less to a depth of at least 30 cm (12 in.) and value of 3 or less and chroma of 1 or less in any remaining layers above the depleted or gleyed matrix. In any sandy material above the depleted or gleyed matrix, at least 70% of the visible soil particles must be masked with organic material, viewed through a 10x or 15x hand lens. Observed without a hand lens, the particles appear to be close to 100% masked.

A13 (Alaska Gleyed)—A mineral layer with a dominant hue of N, 10Y, 5GY, 10GY, 5G, 10G, 5BG, 10BG, 5B, 10B, or 5PB, with value of 4 or more in more than 50% of the matrix. The layer starts within 30 cm (12 in.) of the mineral surface, and is underlain within 1.5 m (60 in.) by soil material with hue 5Y or redder in the same type of parent material.

A14 (Alaska Redox)—A mineral layer that has dominant hue 5Y with chroma of 3 or less, or a gleyed matrix, with 10% or more distinct or prominent redox concentrations occurring as pore linings with value and chroma of 4 or more. The layer occurs within 30 cm (12 in.) of the soil surface.

A15 (Alaska Gleyed Pores)—A mineral layer that has 10% or more hue N, 10Y, 5GY, 10GY, 5G, 10G, 5BG, 10BG, 5B, 10B, or 5PB with value of 4 or more along root channels or other pores and that starts within 30 cm (12 in.) of the soil surface. The matrix has dominant hue of 5Y or redder.

A16 (Coast Prairie Redox)—A layer starting within 15 cm (6 in.) of the soil surface that is at least 10 cm (4 in.) thick and has a matrix chroma of 3 or less with 2% or more distinct or prominent redox concentrations occurring as soft masses and/or pore linings.

(*Continued*)

TABLE 9.1 (*Continued*)

Field Indicators of Hydric Soils of the United States[a]

S. Field indicators for soils with sandy soil materials:

S1 (Sandy Mucky Mineral)—A layer of mucky modified sandy soil mineral 5 cm (2 in.) or more thick starting within 15 cm (6 in.) of the soil surface.

S2 (3 cm Mucky Peat or Peat)—A layer of mucky peat or peat 2.5 cm (1 in.) or more thick with value of 4 or less and chroma of 3 or less, starting within 15 cm (6 in.) of the soil surface and underlain by sandy soil material.

S3 (5 cm Mucky Peat or Peat)—A layer of mucky peat or peat 5 cm (2 in.) or more thick with value of 3 or less and chroma of 2 or less, starting within 15 cm (6 in.) of the soil surface and underlain by sandy soil material.

S4 (Sandy Gleyed Matrix)—A gleyed matrix that occupies 60% or more of a layer starting within 15 cm (6 in.) of the soil surface.

S5 (Sandy Redox)—A layer starting within 15 cm (6 in.) of the soil surface that is at least 10 cm (4 in.) thick and has a matrix with 60% or more chroma of 2 or less and 2% or more distinct or prominent redox concentrations occurring as soft masses and/or pore linings.

S6 (Stripped Matrix)—A layer starting within 15 cm (6 in.) of the soil surface in which iron-manganese oxides and/or organic matter have been stripped from the matrix and the primary base color of the soil material has been exposed. The stripped areas and translocated oxides and/or organic matter form a faintly contrasting pattern of two or more colors with diffuse boundaries. The stripped zones are 10% or more of the volume and are rounded.

S7 (Dark Surface)—A layer 10 cm (4 in.) thick, starting within the upper 15 cm (6 in.) of the soil surface, with a matrix value of 3 or less and chroma of 1 or less. At least 70% of the visible soil particles must be masked with organic material, viewed through a 10x or 15x hand lens. Observed without a hand lens, the particles appear to be close to 100% masked. The matrix color of the layer directly below the dark layer must have the same colors as those described above or any color that has chroma of 2 or less.

S8 (Polyvalue Below Surface)—A layer with value of 3 or less and chroma 1 or less starting within 15 cm (6 in.) of the soil surface. At least 70% of the visible soil particles must be masked with organic material, viewed through a 10x or 15x hand lens. Observed without a hand lens, the particles appear to be close to 100% masked. Directly below this layer, 5% or more of the soil volume has value of 3 or less and chroma of 1 or less, and the remainder of the soil volume has value of 4 or more and chroma of 1 or less to a depth of 30 cm (12 in.) or to the spodic horizon, whichever is less.

S9 (Thin Dark Surface)—A layer 5 cm (2 in.) or more thick within the upper 15 cm (6 in.) of the surface, with value of 3 or less and chroma of 1 or less. At least 70% of the visible soil particles must be masked with organic material, viewed through a 10x or 15x hand lens. Observed without a hand lens, the particles appear to be close to 100% masked. This layer is underlain by a layer or layers with value of 4 or less and chroma of 1 or less to a depth of 30 cm (12 in.) or to the spodic horizon, whichever is less.

S11 (High Chroma Sands)—In coastal zones and dune-and-swale complexes, a layer 5 cm (2 in.) or more thick starting within 10 cm (4 in.) of the surface with chroma of 4 or less and 2% or more distinct or prominent redox concentrations.

F. Field indicators for soils with loamy and clayey soil material:

F1 (Loamy Mucky Mineral)—A layer of mucky modified loamy or clayey soil material 10 cm (4 in.) or more thick starting within 15 cm (6 in.) of the soil surface.

F2 (Loamy Gleyed Matrix)—A gleyed matrix that occupies 60% or more of a layer starting within 30 cm (12 in.) of the soil surface.

F3 (Depleted Matrix)—A layer that has a depleted matrix with 60% or more chroma of 2 or less and that has a minimum thickness of either:

a. 5 cm (2 in.) if the 5 cm is entirely within the upper 15 cm (6 in.) of the soil, or
b. 15 cm (6 in.), starting within 25 cm (10 in.) of the soil surface

F6 (Redox Dark Surface)—A layer that is at least 10 cm (4 in.) thick, is entirely within the upper 30 cm (12 in.) of the mineral soil, and has:

a. matrix value of 3 or less and chroma of 1 or less and 2% or more distinct or prominent redox concentrations as soft masses or pore linings, or
b. matrix value of 3 or less and chroma of 2 or less and 5% or more distinct or prominent redox concentrations as soft masses or pore linings.

(*Continued*)

TABLE 9.1 (*Continued*)

Field Indicators of Hydric Soils of the United States[a]

F7 (Depleted Dark Surface)—Redox depletion, with value of 5 or more and chroma of 2 or less, in a layer that is at least 10 cm (4 in.) thick, is entirely within the upper 30 cm (12 in.) of the mineral soil, and has:
a. matrix value of 3 or less and chroma of 1 or less and 10% or more redox depletions, or
b. matrix value of 3 or less and chroma of 2 or less and 20% or more redox depletions.

F8 (Redox Depressions)—In closed depressions subject to ponding, 5% or more distinct or prominent redox concentrations occurring as soft masses or pore linings in a layer that is 5 cm (2 in.) or more thick and is entirely within the upper 15 cm (6 in.) of the soil.

F9 (Vernal Pools)—In closed depressions subject to ponding, presence of a depleted matrix with 60% or more chroma of 2 or less in a layer 5 cm (2 in.) or more thick entirely within the upper 15 cm (6 in.) of the soil.

F10 (Marl)—A layer of marl with value of 5 or more and chroma less than 2 starting within 10 cm (4 in.) of the soil surface.

F11 (Depleted Ochric)—A layer(s) 10 cm (4 in.) or more thick in which 60% or more of the matrix has value of 4 or more and chroma of 1 or less. The layer is entirely within the upper 25 cm (10 in.) of the soil.

F12 (Iron/Manganese Masses)—On flood plains, a layer 10 cm (4 in.) or more thick with 40% or more chroma of 2 or less and 2% or more distinct or prominent redox concentrations occurring as soft iron-manganese masses with diffuse boundaries. The layer occurs entirely within 30 cm (12 in.) of the soil surface. Iron-manganese masses have value and chroma of 3 or less. Most commonly, they are black. The thickness requirement is waived if the layer is the mineral surface layer.

F13 (Umbric Surface)—In depressions and other concave landforms, a layer 25 cm (10 in.) or more thick, starting within 15 cm (6 in.) of the soil surface, in which the upper 15 cm (6 in.) has value of 3 or less and chroma of 1 or less and in which the lower 10 cm (4 in.) has the same colors as those described above or any other color that has chroma of 2 or less.

F16 (High Plains Depressions)—In closed depressions that are subject to ponding, a mineral soil that has chroma of 1 or less to a depth of at least 35 cm (13.5 in.) and a layer at least 10 cm (4 in.) thick within the upper 35 cm (13.5 in.) of the mineral soil that has either:
a. 1% or more redox concentrations occurring as nodules or concretions, or
b. redox concentrations occurring as nodules or concretions with distinct or prominent corona.

F17 (Delta Ochric)—A layer 10 cm (4 in.) or more thick in which 60% or more of the matrix has value of 4 or more and chroma of 2 or less and there are no redox concentrations. This layer occurs entirely within the upper 30 cm (12 in.) of the soil.

F18 (Reduce Vertic)—In Vertisols and Vertic intergrades, a positive reaction to α-α-dipyridyl that:
a. is the dominant (60% or more) condition of a layer at least 10 cm (4 in.) thick within the upper 30 cm (12 in.) [or at least 5 cm (2 in.)] thick within the upper 15 cm (6 in.) of the mineral or muck soil surface,
b. occurs for at least 7 continuous days and 28 cumulative days, and
c. occurs during a normal or drier season and month (within 16%–84% of probable precipitation).

F19 (Piedmont Flood Plain Soils)—On active flood plains, a mineral layer at least 15 cm (6 in.) thick starting within 25 cm (10 in.) of the soil surface, with a matrix (60% or more of the volume) chroma of less than 4% and 20% or more distinct or prominent redox concentrations occurring as soft masses or pore linings.

F20 (Anomalous Bright Loamy Soils)—Soils within 200 m (656 ft.) of estuarine marshes or water and within 1 m (3.28 ft.) of mean high water, a mineral layer at least 10 cm (4 in.) thick, starting within 20 cm (8 in.) of the soil surface, with a matrix (60% or more of the volume) chroma of less than 5% and 10% or more distinct or prominent redox concentrations occurring as soft masses or pore linings and/or depletions.

F21 Red Parent Material. A layer derived from red parent materials that is at least 10 cm (4 in.) thick, starting within 25 cm (10 in.) of the soil surface with a hue of 7.5YR or redder. The matrix has a value and chroma greater than 2 and less than or equal to 4. The layer must contain 10% or more depletions and/or distinct or prominent redox concentrations occurring as soft masses or pore linings. Redox depletions should differ in color by having:
a. value one or more higher and chroma one or more lower than the matrix, or
b. value of 4 or more and chroma of 2 or less.

(*Continued*)

TABLE 9.1 (*Continued*)

Field Indicators of Hydric Soils of the United States[a]

F22 (Very Shallow Dark Surface)—In depressions and flood plains subject to frequent ponding and/or flooding, one of the following: a. If bedrock occurs between 15 cm (6 in.) and 25 cm (10 in.), a layer at least 15 cm (6 in.) thick starting within 10 cm (4 in.) of the soil surface with value 2.5 or less and chroma 1 or less, and the remaining soil to bedrock must have the same colors as above or any other color that has a chroma 2 or less. b. If bedrock occurs within 15 cm (6 in.), more than half of the soil thickness must have value 2.5 or less and chroma 1 or less, and the remaining soil to bedrock must have the same colors as above or any other color that has a chroma 2 or less.

[a] Field indicators of hydric soils have been approved for use in each land resource region (USDA NRCS 2010; see Table 9.4.).

the range of estimated properties for a soil is within any portion of any of the hydric soil criteria, that soil will appear on hydric soils lists. Conversely, the absence of a soil from the hydric soil list does not preclude the presence of hydric soils.

National Wetland Inventory Maps

Also available for offsite examination are NWI maps produced by the USFWS. NWI maps contain wetland delineations as defined in "Classification of Wetlands and Deepwater Habitats of the United States" (Cowardin et al. 1979) at a scale of 1:24,000. The NWI maps were produced by interpreting high-altitude photography, usually at a scale of 1:40,000–1:80,000. The NWI have three limitations for wetland delineation. First, the definition of wetlands used to produce the NWI maps is not the same as the definitions used to delineate jurisdictional wetlands. Jurisdictional wetlands are determined based on the three factors of soils, hydrology, and vegetation whereas NWI wetland maps may have delineations based on only one factor and often fail to delineate cropped fields and borderline wetlands. Second, many NWI maps were produced from poor-quality aerial photography. Finally, scale limitations do not allow for delineation of areas less than about 1.6 hectares.

Topographic Maps

Another source of information is the topographic quadrangle series of maps produced by USGS. These maps contain topographic features including swamp and marsh symbols at a scale of 1:24,000 and may be useful as a source of offsite wetland information. Limitations of these maps for wetland delineation include the following points. First, not all areas with marsh and swamp symbols are wetlands. Conversely, there are areas of wetlands that lack marsh and swamp symbols. Second, the quality of the topographic maps varies from quadrangle to quadrangle and within any given quadrangle; however, the degree of field verification is indicated on the legend for each map. Finally, the scale limitation is the same as for the NWI maps.

Federal Emergency Management Agency Maps

Another source of information is the topographic quadrangle series of maps produced by the FEMA. These maps contain delineations of areas that FEMA has determined are flood prone at a scale of 1:24,000. The limitations of FEMA maps for wetland delineation include the following. First, flood-prone areas delineated contain many areas of uplands flooded at a frequency ranging from once every 1 to –500 years. Although many areas of wetlands will be within areas delineated as flood-prone areas, there will also be many areas of uplands. Second, saturated wetlands and many depressional wetlands are not identified on these maps. Third, flood-prone maps do not provide information related to the duration of flooding, which is essential to the hydric soil definition. Finally, the scale limitation is approximately the same as for the NWI maps and the USGS topography quadrangle maps.

Because of the limitations listed above, onsite investigation is usually necessary to decide if hydric soils occur and to determine the exact location and extent of hydric soils. However, valuable insight can be gained by reviewing these sources of information before attempting hydric soil delineations, reducing the time needed to locate and delineate hydric soils.

Detailed Examination and Delineation Procedures

Landform Recognition

A landscape is the land surface that an eye can comprehend in a single view (Tuttle 1975; U.S. Department of Agriculture 1993). Most frequently it is a collection of landforms. Landforms are physical, recognizable forms or features on the earth's surface that have characteristic shapes produced by natural processes. Hydric soils occur on landforms (U.S. Department of Agriculture 1993) that include, but are not limited to, backswamps, bogs, depressions, estuaries, fens, interdunes, marshes, flats, floodplains, muskegs, oxbows, playas, pocosins, potholes, seep slopes, sloughs, and swamps (Figure 9.1). One of the most important factors in hydric soil determination and delineation is landform recognition.

Hydric soils develop when water saturates the soil or collects on the soil surface, oxygen is removed and the soil becomes anaerobic. A concave surface frequently augmented by

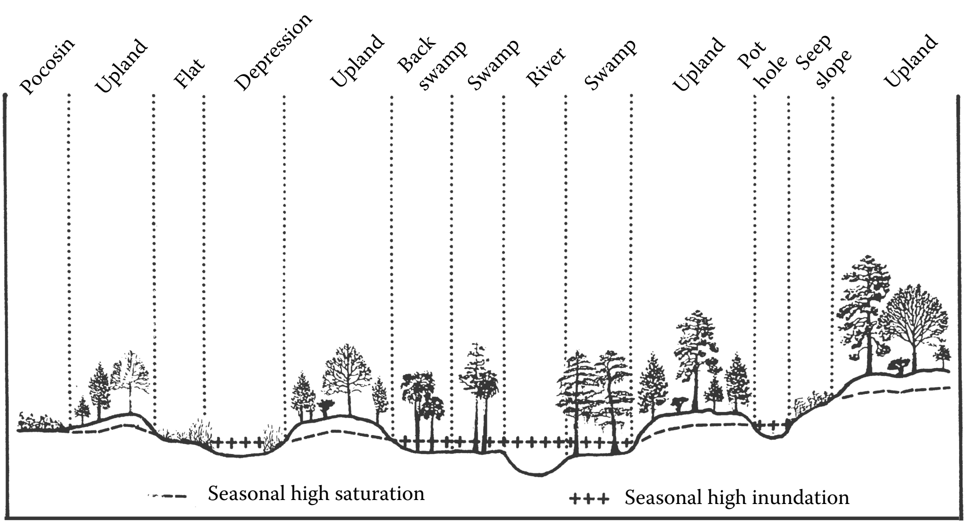


FIGURE 9.1
Idealized landscape depicting uplands and the hydric soil landforms pocosin, flat, depression, back swamp, swamp, pot hole, and seep slope. Note that each hydric soil area begins at a slightly concave slope break, although not all of each hydric soil area expresses concavity throughout the landform (seep slope). Vertical scale is exaggerated.

slower percolating subsurface soil horizons allows this process to occur. Field indicators normally begin to appear at this concave slope break and continue throughout the extent of the wetland even though concavity may not exist throughout the wetland (see Figure 9.1). The concave slope break may be very subtle, but it will be present in almost all natural landscapes. Wetland delineators need to become very familiar with the landscapes and hydrology of their areas in order to recognize the often very subtle slope break. They need to anticipate where inundated or saturated soils are likely to occur. Water is the driving force behind the development of hydric soils and wetlands. Hydrology of the landscape must be understood prior to making hydric soil determinations and delineating wetlands.

Field Indicators of Hydric Soils

Field indicators are formed predominantly by accumulation, loss, or transformation of iron, manganese, sulfur, or carbon compounds. The presence of hydrogen sulfide (H_2S) which has a "rotten egg odor" is a strong field indicator of a hydric soil, but is found in only the wettest sites containing sulfur. While field indicators related to iron/manganese (Fe/Mn) depletions or concentrations are the most common, they cannot form in soils with parent materials that contain very low amounts of Fe/Mn. Soils formed in such materials may have low chroma colors (two or less) that are not related to saturation and reduction. For these soils, features related to accumulations of organic carbon are most commonly used.

Field indicators of hydric soils are routinely used in conjunction with the hydric soils definition to confirm the presence or absence of a hydric soil. The publication of the Field Indicators (USDA NRCS 2010) is the current guide that should be applied to identify and delineate hydric soils in the field. NTCHS is responsible for revising and maintaining the field indicators and any updates to the Field Indicators can be found in the errata posted on the NTCHS web page. Field indicators currently approved for identifying and delineating hydric soils are given in Table 9.1; photographs of each are provided in USDA NRCS (2010).

All of the Field Indicators are formatted in the same structured presentation: (1) letter/number symbol, (2) short descriptive name, (3) geographic region of application, (4) technical requirements, and (5) user notes. For example, each field indicator has a symbol A1, S2, F3, etc. The letter designation identifies what group of USDA soil texture the field indicator applies. Field indicators with an "A" designation can be used for soils without regard to texture. Field indicators with and "S" designation can only be used for soils that have a USDA texture of loamy fine sand and coarser. Field indicators with an "F" designation can only be used for soils that have a USDA texture of loamy very fine sand and finer. All field indicators preceded with "T" are to be used for testing. The symbol is followed by a short descriptive name, example Hydrogen Sulfide, Sandy Redox, or Depleted Matrix. The third part of each field indicator describes the geographic area where the field indicator is known to occur and can be applied. The area or regions of application are typically identified by land resource region (LRR). LRRs are USDA ecoregion classifications (Figure 9.2). In some cases, a field indicator is restricted to subregions within an LRR which are major land resource area (MLRA). Additional information regarding LRRs and MLRAs can be found in *Land Resource Regions and Major Land Resource Areas of the United States, the Caribbean, and the Pacific Basin*—U.S. Department of Agriculture Handbook 296 (2006). The fourth part describes the specific requirements of each field indicator. Changes to any part of the field indicator can only be made by the NTCHS. The last part of the field indicator is the "user notes." User notes are information about the field indicator that maybe helpful in identifying a field indicator. User notes are described for each field indicator, but user notes are the single part of a field indicator that can be supplemented by working groups

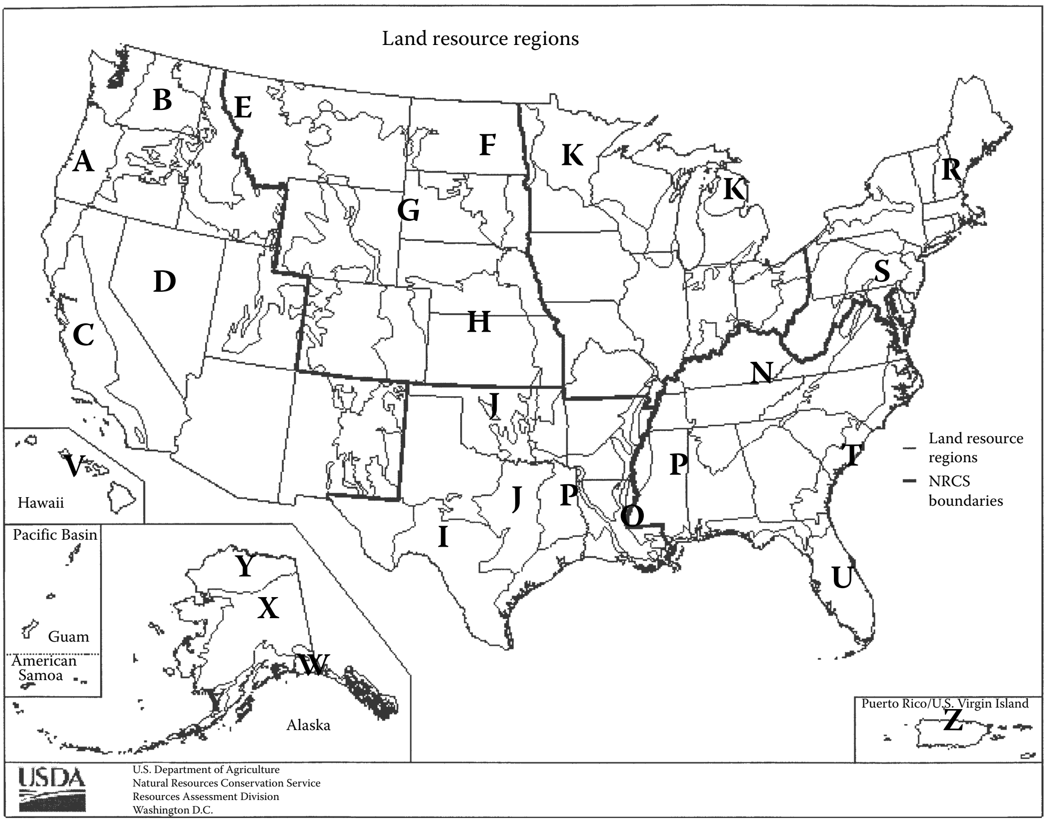


FIGURE 9.2
Land Resource Regions of the United States and Puerto Rico. (Adapted from U.S. Department of Agriculture, Soil Conservation Service. 1981. *Land Resource Regions and Major Land Resource Areas of the United States*. USDA–SCS Agricultural Handbook 296. U.S. Govt. Printing Office. Washington, DC.)

of soil scientists within a region or state or even an individual. However, any additions to user notes cannot make the field indicator more inclusive or restrictive, change the geographic area of application, or requirements or wording of the field indicator. An example of the various parts of field indicator F2 (Loamy Gleyed Matrix) is described in Table 9.2.

"F2. Loamy Gleyed Matrix. For use in all LRRs, except for W, X, and Y. A gleyed matrix that occupies 60% or more of a layer starting within 30 cm (12 in.) of the soil surface.

User Notes: Gley colors are not synonymous with gray colors. They are the colors on the gley color pages of the Munsell color book (Xrite 2009). They have hue of N, 10Y, 5GY, 10GY, 5G, 10G, 5BG, 10BG, 5B, 10B, or 5PB and value of 4 or more. The gleyed matrix only has to be present within 30 cm (12 in.) of the surface. Soils with gleyed matrices are saturated for periods of a significant duration; as a result, there is no thickness requirement for the layer."

The requirements for each field indicator are specific for that field indicator; the following statement applies to all field indicators unless specifically excluded in the preamble for each group of field indicators: "All mineral layers above any of the layers meeting the requirements of any indicator(s), except for indicator A16, S6, F8, F12, F19, and F20 have a dominant chroma of 2 or less, or the thickness of the layer(s) with a dominant chroma of more than 2 is less than 15 cm (6 in.). Except for indicator F16 nodules and concretions are not considered to be redox concentrations."

TABLE 9.2

Field Indicators of Hydric Soils of the United States

Indicator Symbol	Indicator Name	Region of Application	Indicator	User Notes
F2	Loamy Gleyed Matrix	For use in all LRRs, except for W, X, and Y	A gleyed matrix that occupies 60% or more of a layer starting within 30 cm (12 in.) of the soil surface	Gley colors are not synonymous with gray colors. They are the colors on the gley color pages of the Munsell color book (Xrite 2009). They have hue of N, 10Y, 5GY, 10GY, 5G, 10G, 5BG, 10BG, 5B, 10B, or 5PB and value of 4 or more. The gleyed matrix only has to be present within 30 cm (12 in.) of the surface. Soils with gleyed matrices are saturated for periods of a significant duration; as a result, there is no thickness requirement for the layer.

The list of Field Indicators is not static. Changes are anticipated as new knowledge of morphological, physical, chemical, and mineralogical soil properties accumulates. Revisions and additions will continue as we gain a better understanding of the relationships between the development of recognizable soil properties and anaerobic soil conditions. Field indicators that NTCHS has identified for testing are given in Table 9.3. Comments regarding the test field indicators and field observations of hydric soil conditions that cannot be documented using the presently recognized field indicators are welcome; however, any modifications must be approved by NTCHS. The procedure for suggesting changes and commenting on field indicators can be found in the Field Indicators (USDA NRCS 2010). Many of these test field indicators are known to provide reliable guidelines for hydric soil delineation.

A minimal number of terms (Table 9.4) must be defined correctly to interpret Tables 9.1 and 9.3. Some of these terms and definitions are specific to the identification of field indicators and may vary from the definitions in other references. To apply field indicators

TABLE 9.3

Test Field Indicators of Hydric Soils of the United States[a]

TA. Test field indicators for all soils regardless of texture:

TA4 (Alaska Color Change)—A mineral layer 10 cm (4 in.) or more thick, starting within 30 cm (12 in.) of the surface, that has a matrix value of 4 or more and chroma of 2 or less and that within 30 minutes becomes redder by one or more Munsell unit in hue and/or chroma when exposed to air.

TA5 (Alaska Alpine Swales)—On concave landforms, the presence of a surface mineral layer 10 cm (4 in.) or more thick having hue of 10YR or yellower, value 2.5 or less, and chroma of 2 or less. The dark layer is at least twice as thick as the mineral surface layer of soils in the adjacent convex micro-positions.

TA6 (Mesic Spodic)—A layer 5 cm (2 in.) or more thick, starting within 15 cm (6 in.) of the mineral soil surface, that has value of 3 or less and chroma of 2 or less and is underlain by either:

a. a layer(s) 8 cm (3 in.) or more thick occurring within 30 cm (12 in.) of the mineral soil surface, having value and chroma of 3 or less, and showing evidence of spodic development, or
b. a layer(s) 5 cm (2 in.) or more thick occurring within 30 cm (12 in.) of the mineral soil surface, having value of 4 or more and chroma of 2 or less, and directly underlain by a layer(s) 8 cm (3 in.) or more thick having value and chroma of 3 or less and showing evidence of spodic development.

TABLE 9.4

Definition of Terms[a]

Depleted Matrix—A depleted matrix refers to the volume of a soil horizon or subhorizon from which iron has been removed or transformed by processes of reduction and translocation to create colors of low chroma and high value. A, E, and calcic horizons may have low chromas and high values and may therefore be mistaken for a depleted matrix; however, they are excluded from the concept of depleted matrix unless the soil has common or many distinct or prominent redox concentrations occurring as soft masses or pore linings. In some areas, the depleted matrix may change color upon exposure to air (reduced Matrix); this phenomenon is included in the concept of depleted matrix. The following combinations of value and chroma identify a depleted matrix:

1. Matrix value of 5 or more and chroma of 1 or less with or without redox concentrations occurring as soft masses and/or pore linings; or
2. Matrix value of 6 or more and chroma of 2 or less with or without redox concentrations as soft masses and/or pore linings; or
3. Matrix value of 4 or 5 and chroma of 2 and has 2% or more distinct or prominent redox concentrations as soft masses and/or pore linings; or
4. Matrix value of 4 and chroma of 1 and has 2% or more distinct or prominent redox concentrations as soft masses and/or pore linings.
5. Diffuse Boundary—Used to describe redoximorphic features that grade gradually from one color to another. The color grade is commonly more than 2 mm wide. Clear is used to describe boundary color gradations intermediate between sharp and diffuse.

Distinct—Readily seen but contrast only moderately with the color to which compared; a class of contrast intermediate between faint and prominent. In the same hue or a difference in hue of one color chart (e.g., 10YR to 7.5YR or 10YR to 2.5Y), a change of 2 or 3 units in chroma and/or a change of 3 units of value, or a change of 2 or 3 units of value and a change of 1 or 2 units of chroma, or a change of 1 unit of value and 2 units of chroma. With a change of 2 color charts of hues (e.g., 10YR to 5Y or 10YR to 5YR), a change of 0 to 2 units of value and/or a change of 0 to 2 units of chroma is distinct.

Faint—Evident only on close examination. In the same hue or 1 hue change (e.g., 10YR to 7.5YR or 10YR to 2.5Y) a change of 1 unit in chroma, or 1 to 2 units in value, or 1 unit of chroma and 1 unit of value.

Gilgai—A type of microrelief produced by expansion and contraction of soils that results in enclosed microbasins and microknolls.

Glauconitic—A mineral aggregate that contains micaceous mineral resulting in a characteristic green color (e.g., glauconitic shale or clay).

Gleyed Matrix—Soils with a gleyed matrix have the following combinations of hue, value, and chroma, and the soils are not glauconitic:

1. 10Y, 5GY, 10GY, 10G, 5BG, 10BG, 5B, 10B, or 5PB with value 4 or more and chroma is 1; or
2. 5G with value 4 or more and chroma is 1 or 2; or
3. N with value 4 or more; or
4. (For testing only) 5Y, value 4, and chroma 1.

In some places the gleyed matrix may change color upon exposure to air (reduced matrix). This phenomenon is included in the concept of gleyed matrix.

Hemic—See Mucky Peat.

Histic Epipedon—A thick (20–60 cm, or 8–24 in.) organic soil horizon that is saturated with water at some period of the year unless artificially drained and that is at or near the surface of a mineral soil.

Hydric Soil Definition (1994)—A soil that formed under conditions of saturation, flooding, or ponding long enough during the growing season to develop anaerobic conditions in the upper part.

Loamy and Clayey Soil Material—Refers to those soil materials with a USDA texture of loamy very fine sand and finer.

Masked—Through redoximorphic processes, the color of soil particles is hidden by organic material, silicate clay, iron, aluminum, or some combination of these.

Muck—Sapric organic soil material in which virtually all of the organic material is so decomposed that identification of plant forms is not possible. Bulk density is normally 0.2 or more. Muck has less than one-sixth fibers after rubbing, and its sodium pyrophosphate solution extract color has lower value and chroma than 5/1, 6/2, and 7/3.

(*Continued*)

TABLE 9.4 (*Continued*)

Definition of Terms[a]

Mucky Modified Texture—A USDA soil texture modifier (e.g., mucky sand). Mucky-modified mineral soil material that has 0% clay has between 5% and 12% organic carbon. Mucky-modified mineral soil material that has 60% clay has between 12% and 18% organic carbon. Soils with an intermediate amount of clay have intermediate amounts of organic carbon. Where the organic component is peat (fibric material) or mucky peat (hemic material), mucky mineral soil material does not occur.
Mucky Peat—Hemic organic material, which is characterized by decomposition that is intermediate between that of fibric material and that of sapric material. Bulk density is normally between 0.1 and 0.2 g/cm^3. Mucky peat does not meet the fiber content (after rubbing) or sodium pyrophosphate solution extract color requirements for either fibric or sapric soil material.
Organic Soil Material—Soil material that is saturated with water for long periods or artificially drained, and excluding live roots, has 18% or more organic carbon with 60% or more clay or 12% or more organic carbon with 0% clay. Soils with an intermediate amount of clay have an intermediate amount of organic carbon. If the soil is never saturated for more than a few days, it contains 20% or more organic carbon. Organic soil material includes muck, mucky peat, and peat (Chapter 8).
Peat—Fibric organic soil material. The plant forms can be identified in virtually all of the organic material. Bulk density is normally <0.1. Peat has three-fourths or more fibers after rubbing, or it has two-fifths or more fibers after rubbing and has sodium pyrophosphate solution extract color of 7/1, 7/2, 8/2, or 8/3.
Prominent—Soils contrasting strongly with the color to which they are compared. In the same hue or a 1 hue change (e.g., 10YR to 2.5Y or 10YR to 7.5YR), a change of 4 units in chroma and/or 4 units in value. With a change of 2 hues (e.g., 10YR to 5Y or 10YR to 5YR), a change of 3 or more units of value and/or a change of 3 or more units of chroma is prominent.
Sandy Soil Material—Refers to those soil materials with a USDA texture of loamy fine sand and coarser.
Sapric—See Muck.
Sharp Boundary—Used to describe redoximorphic features that grade sharply from one color to another. The color grade is commonly less than 0.1 mm wide. Clear is used to describe boundary color gradations intermediate between sharp and diffuse.

[a] These definitions are needed to understand certain terms used in Tables 9.1 and 9.3.

properly, a basic knowledge of soil science, soil–landscape relationships, and soil survey procedures is also necessary. Many field indicators are landform specific. Professional soil or wetland scientists familiar with local conditions are best equipped to make an onsite hydric soil determination.

Combining Field Indicators of Hydric Soils

It is permissible to combine certain field indicators if all requirements of the individual field indicator are met except thickness (see hydric soil technical note 4, http://soils.usda.gov/use/hydric/ntchs/tech_notes/index.html). The most restrictive requirements for thickness of layers in any field indicators used must be met. Not all field indicators are possible candidates for combination. For example, field indicator F2 (Loamy Gleyed Matrix) has no thickness requirement, so a site would either meet the requirements of this field indicator or it would not. Table 9.5 lists the field indicators that are the most likely candidates for combining in the region.

Table 9.6 presents an example of a soil in which a combination of layers meets the requirements for field indicators F6 (Redox Dark Surface) and F3 (Depleted Matrix). The second layer meets the morphological characteristics of F6 and the third layer meets the morphological characteristics of F3, but neither meets the thickness requirement for its respective field indicator. However, the combined thickness of the second and third layers meets the more restrictive conditions of thickness for F3 [i.e., 15 cm (6 in.) starting within 25 cm

TABLE 9.5

Minimum Thickness Requirements for Commonly Combined Field Indicators of Hydric Soils

Field Indicator	Thickness Requirement
S5 (Sandy Redox)	10 cm (4 in.) thick starting within 15 cm (6 in.) of the soil surface
S7 (Dark Surface)	10 cm (4 in.) thick starting within 15 cm (6 in.) of the soil surface
F1 (Loamy Mucky Mineral)	10 cm (4 in.) thick starting within 15 cm (6 in.) of the soil surface
F3 (Depleted Matrix)	15 cm (6 in.) thick starting within 25 cm (10 in.) of the soil surface
F6 (Redox Dark Surface)	10 cm (4 in.) thick entirely within the upper 30 cm (12 in.)
F7 (Depleted Dark Surface)	10 cm (4 in.) thick entirely within the upper 30 cm (12 in.)

TABLE 9.6

Example of a Soil That Is Hydric Based on a Combination of Field Indicators F6 and F3

		Redox Concentrations			
Depth (cm)	**Matrix Color**	**Color**	**Abundance**	**Contrast**	**Texture**
0–8	10YR 2/1	–	–	–	Loamy/clayey
8–15	10YR 3/1	7.5YR 5/6	3%	Prominent	Loamy/clayey
15–25	10YR 5/2	7.5YR 5/6	5%	Prominent	Loamy/clayey
25–36	2.5Y 4/2	–	–	–	Loamy/clayey

TABLE 9.7

Example of a Soil That Is Hydric Based on a Combination of Field Indicators F6 and F3

		Redox Concentrations			
Depth (cm)	**Matrix Color**	**Color**	**Abundance**	**Contrast**	**Texture**
0–8	10YR 3/1	10YR 5/6	3%	Prominent	Loamy/clayey
8–15	10YR 4/1	10YR 5/6	3%	Prominent	Sandy
15–25	10YR 4/1	–	–	–	Loamy/clayey

(10 in.) of the soil surface]. Therefore, the soil is considered to be hydric based on the combination of field indicators.

Another common situation in which it is appropriate to combine the characteristics of field indicators is when stratified textures of sandy (i.e., loamy fine sand and coarser) and loamy/clayey (i.e., loamy very fine sand and finer) material occur in the upper 30 cm (12 in.) of the soil. For example, the soil shown in Table 9.7 is hydric based on a combination of field indicators F6 (Redox Dark Surface) and S5 (Sandy Redox). This soil meets the morphological characteristics of F6 in the first layer and S5 in the second layer, but neither layer by itself meets the thickness requirement for its respective field indicator. However, the combined thickness of the two layers [15 cm (6 in.)] meets the more restrictive thickness requirement of either field indicator [10 cm (4 in.)].

Field Indicators of Hydric Soils for Delineation and Identification

Field Indicators were originally developed to delineate the boundary between wetlands and uplands. Typically the soils near the boundary are saturated near the soil surface for only a few weeks a year. The repeated saturation and drying of the soil over time often develop a strong expression of redox features. However, some soils that remain saturated nearly all year do not develop the same obvious redox features. This lack of redox development does not mean that field indicators are not present or that these areas do not meet the definition of a hydric soil and are therefore hydric.

Table 9.8 differentiates those field indicators used primarily for hydric soil delineation and those used primarily for identification. Those field indicators identified primarily for the identification of hydric soils usually occur in the wettest portion of wetlands that are normally saturated or inundated for much of most years, and those field indicators identified primarily for delineation occur at the much drier wetland boundary.

Field indicators A1 (Histosols or Histel), A2 (Histic Epipedon), and A3 (Black Histic) are not normally used to identify the delineation boundary of hydric soils except possibly in Alaska (LRRs W, X, and Y). Other field indicators with organic soil material (e.g., A8, A9, and A10) are used more often to delineate hydric soils. If field indicator A1 is used to identify hydric soils, organic soil material and Histosol or Histel requirements contained in *Soil Taxonomy* must be met (U.S. Department of Agriculture, Soil Survey Staff 1999, pp. 20–21, 86–113, and 473–487). If field indicator A2 is used to identify hydric soils, all the requirements contained in *Soil Taxonomy* must be met (U.S. Department of Agriculture, Soil Survey Staff 1999, pp. 22–23). Unlike field indicators A1 and A2, no taxonomic requirements exist for A3. Field indicator A3 identifies those Histic Epipedons that are always wet in natural conditions.

Field indicators A4 (Hydrogen Sulfide), S4 (Sandy Gleyed Matrix), and F2 (Loamy Gleyed Matrix) are not normally used to identify the delineation boundary of hydric soils. Presence of "rotten egg" odor for A4 and the gleying for S4 and F2 indicates the soils are reduced for much of each year and would therefore identify only the wetlands saturated or inundated for very long periods. These three field indicators normally occur inside the delineation line established by the delineation field indicators.

Field indicator A5 (Stratified Layers) is routinely used to delineate hydric soils on floodplains and some flats. Soils on the non-hydric side of delineations are stratified, but the chroma in one or more layers is 3 or higher.

Field indicator A6 (Organic Bodies) is routinely used to delineate hydric soils dominantly on flats of the southern United States and Puerto Rico. Soils on the non-hydric side of delineations usually have organic accreted areas, but these bodies lack the required amount of organic carbon.

Field indicators A7 (5 cm Mucky Mineral), A8 (Muck Presence), A9 (1 cm Muck), A10 (2 cm Muck), S1 (Sandy Mucky Mineral), S2 (3 cm Mucky Peat or Peat), S3 (5 cm Mucky Peat or Peat), and F1 (Loamy Mucky Mineral) are routinely used to delineate hydric soils throughout various regions of the United States and Puerto Rico. Soils on the non-hydric side of delineations usually have surface layers that lack the required amount of organic carbon.

Field indicator A11 (Depleted Below Dark Surface) is used to delineate hydric soils nationwide. Field indicator A12 (Thick Dark Surface) is not normally used to delineate hydric soils. This field indicator normally occurs inside the delineation line established by field indicator A11, F6 (Redox Dark Surface), or F3 (Depleted Matrix).

Field indicators S10 (Alaska Gleyed), A13 (Alaska Gleyed), A14 (Alaska Redox), and A15 (Alaska Gleyed Pores) are used for delineation purposes in Alaska. Soils on the non-hydric side of delineations usually lack chroma 2 or less within the required depths or lack the required amounts and kinds of redox features. A16 (Coast Prairie Redox) is used to delineate hydric soils on depressions and intermounds on the Lissie Formation in Texas.

TABLE 9.8

Field Indicators of Hydric Soils Used Primarily for Delineation and Identification

Field Indicator		Type of Field Indicator
AI	Histosol or Histel	Identification
A2	Histic Epipedon	Identification
A3	Black Histic	Identification
A4	Hydrogen Sulfide	Identification
A5	Stratified Layers	Delineation
A6	Organic Bodies	Delineation
A7	5 cm Mucky Mineral	Delineation
A8	Muck Presence	Delineation
A9	1 cm Muck	Delineation
A10	2 cm Muck	Delineation
A11	Depleted Below Dark Surface	Delineation
A12	Thick Dark Surface	Identification
A13	Alaska Gleyed	Delineation
A14	Alaska Redox	Delineation
A15	Alaska Gleyed Pores	Delineation
A16	Coast Prairie Redox	Delineation
S1	Sandy Mucky Mineral	Delineation
S2	3 cm Mucky Peat or Peat	Delineation
S3	5 cm Mucky Peat or Peat	Delineation
S4	Sandy Gleyed Matrix	Identification
S5	Sandy Redox	Delineation
S6	Stripped Matrix	Delineation
S7	Dark Surface	Delineation
S8	Polyvalue Below Surface	Delineation
S9	Thin Dark Surface	Delineation
F1	Loamy Mucky Mineral	Delineation
F2	Loamy Gleyed Matrix	Identification
F3	Depleted Matrix	Delineation
F6	Redox Dark Surface	Delineation
F7	Depleted Dark Surface	Delineation
F8	Redox Depressions	Delineation
F9	Vernal Pools	Delineation
F10	Marl	Delineation
F11	Depleted Ochric	Delineation
F12	Iron/Manganese Masses	Delineation
F13	Umbric Surface	Delineation
F16	High Plains Depressions	Delineation
F17	Delta Ochric	Delineation
F18	Reduced Vertic	Delineation
F19	Piedmont Flood Plain Soils	Delineation
F20	Anomalous Bright Loamy Soils	Delineation
F21	Red Parent Material	Delineation
F22	Very Shallow Dark Surface	Delineation

S5 (Sandy Redox), S6 (Stripped Matrix), S7 (Dark Surface), S8 (Polyvalue Below Surface), and S9 (Thin Dark Surface) are routinely used to delineate hydric soils throughout various regions of the United States and Puerto Rico. Soils on the non-hydric side of delineations usually lack chroma 2 or less within 6 in. of the surface (S5), have a layer that meets all the requirements of a stripped matrix except depth (S6), or the surface layer has a salt-and-pepper appearance (S7).

Field indicators S8 (Polyvalue Below Surface) and S9 (Thin Dark Surface) are not normally used to identify the delineation boundary of hydric soils. These two field indicators normally occur inside the delineation line established by field indicators S5, S6, and/or S7.

Field indicators F3 (Depleted Matrix), F6 (Redox Dark Surface), F7 (Depleted Dark Surface), and F13 (Umbric Surface) are routinely used to delineate hydric soils throughout various regions of the United States and Puerto Rico. Soils on the non-hydric side of delineations usually lack chroma 2 or less within the required depths or lack the required amounts and kinds of redox features (F3, F6, and F7) or the surface layer is too thin or not dark enough (F13).

Field indicators F8 (Redox Depressions), F9 (Vernal Pools), and F16 (High Plains Depressions) are used to delineate hydric soils that occur in closed depressions subject to ponding throughout various regions of the United States and Puerto Rico. Soils on the non-hydric side of delineations usually lack any redox features within the required depths.

Field indicator F10 (Marl) is used to delineate hydric soils in southern Florida and in the near shore regions of the Great Lakes. Soils on the non-hydric side of delineations may meet all the requirements of marl, but the chroma is 2 or more or they are dry Histosols (Folists).

Field indicators F11 (Depleted Ochric), F12 (Iron/Manganese Masses), and F17 (Delta Ochric) are used to delineate hydric soils that occur on floodplains that frequently flood predominantly in the southern United States. Soils on the non-hydric side of delineations usually lack any redox features within the required depths.

F18 (reduced Vertic) is used to delineate hydric soils in areas where Vertisols and Vertic intergrades occur in Texas. F19 (Piedmont Flood Plain Soils) is used for delineation purposes on flood plains subject to Piedmont deposition. F20 (Anomalous Bright Loamy Soils) is used to delineate areas of hydric soils adjacent to estuarine areas in the Mid-Atlantic States. F21 (Red Parent Material) is used to identify soils with red parent material in part of the Mid-Atlantic States. F22 (Very Shallow Dark Surface) is used to delineate hydric soils on depressions and flood plains in parts of Florida.

Regional Field Indicators of Hydric Soils

Why Regional Field Indicators?

During 1994 and 1995, a national team of soil scientists and other wetland scientists representing the USDA, NRCS, EPA, the USFWS, and the USACE, universities, and the private sector tested proposed field indicators nationwide (Chapter 2). Quickly, during the review process this team realized: (1) some field indicators, such as A1 (Histosol), expressed a maximum of anaerobiosis and identified but did not delineate hydric soils nationwide, (2) some redox-process-driven field indicators, such as S6 (Stripped Matrix) and F3 (Depleted Matrix), identified and delineated hydric soils virtually nationwide, and (3) some field

TABLE 9.9

For Use Field Indicators of Hydric Soils by Land Resource Region (LRR)

LRR	Field Indicators
A	A1, A2, A3, A4, A11, A12, S1, S4, S5, S6, F1 (except MLRA 1), F2, F3, F6, F7, F8
B	A1, A2, A3, A4, A11, A12, S1, S4, S5, S6, F1, F2, F3, F6, F7, F8, F9
C	A1, A2, A3, A4, A5, A11, A12, S1, S4, S5, S6, F1, F2, F3, F6, F7, F8, F9
D	A1, A2, A3, A4, A9, A11, A12, S1, S4, S5, S6, F1, F2, F3, F6, F7, F8, F9
E	A1, A2, A3, A4, A11, A12, S1, S4, S5, S6, F1, F2, F3, F6, F7, F8
F	A1, A2, A3, A4, A5, A9, A11, A12, S1, S3, S4, S5, S6, F1, F2, F3, F6, F7, F8
G	A1, A2, A3, A4, A9, A11, A12, S1, S2, S4, S5, S6, F1, F2, F3, F6, F7, F8
H	A1, A2, A3, A4, A9, A11, A12, S1, S2, S4, S5, S6, F1, F2, F3, F6, F7, F8, F16 (MLRAs 72 and 73)
I	A1, A2, A3, A4, A11, A12, S1, S4, S5, S6, F1, F2, F3, F6, F7, F8
J	A1, A2, A3, A4, A11, A12, S1, S4, S5, S6, F1, F2, F3, F6, F7, F8
K	A1, A2, A3, A4, A5, A11, A12, S1, S4, S5, S6, S7, S11, F1, F2, F3, F6, F7, F8, F10
L	A1, A2, A3, A4, A5, A11, A12, S1, S4, S5, S6, S7, S11, F1, F2, F3, F6, F7, F8, F10
M	A1, A2, A3, A4, A5, A10, A11, A12, S1, S3, S4, S5, S6, S7, F1, F2, F3, F6, F7, F8
N	A1, A2, A3, A4, A5, A10, A11, A12, S1, S4, S5, S6, S7, F2, F3, F6, F7, F8, F12, F13 (MLRA 138), F21 (MLRA 127)
O	A1, A2, A3, A4, A5, A11, A12, S1, S4, S5, S6, F1, F2, F3, F6, F7, F8, F12
P	A1, A2, A3, A4, A5, A6 (except MLRA 136), A7 (except MLRA 136), A9 (except MLRA 136), A11, A12, S4, S5, S6, S7, F2, F3, F8, F12, F13, F22 (MLRA 138)
Q	A1, A2, A3, A4, A8, A11, A12, S1, S4, S7, F2, F3, F6, F7, F8
R	A1, A2, A3, A4, A5, A11, A12, S1, S4, S5, S6, S7, S8, S9, F2, F3, F6, F7, F8
S	A1, A2, A3, A4, A5, A11, A12, S1, S4, S5, S6, S7, S8, S9, F2, F3, F6, F7, F8, F19 (MLRAs 148 and 149A), F29 (MLRA 149A), F21 (MLRA 145, 147 and 148)
T	A1, A2, A3, A4, A5, A6, A7, A9, A11, A12, A16 (MLRA 150A), S4, S5, S6, S7, S8, S9, F2, F3, F6, F8, F11 (MLRA 151), F12, F13, F17 (MLRA 151), F18 (MLRA 150), F20 (MLRAs 153C and 153D), F22 (West Florida Portion of MLRA 152A)
U	A1, A2, A3, A4, A5, A6, A7, A8, A11, A12, S4, S5, S6, S7, S8, S9, F2, F3, F6, F7, F8, F10, F13, F22 (MLRA 154)
V	A1, A2, A3, A4, A5, A8, A11, A12, S1, S4, S7, F2, F3, F6, F7, F8
W	A1, A2, A3, A4, A12, A13, A14, A15
X	A1, A2, A3, A4, A12, A13, A14, A15
Y	A1, A2, A3, A4, A12, A13, A14, A15
Z	A1, A2, A3, A4, A5, A6, A7, A8, A11, A12, S4, S5, S6, S7, F2, F3, F6, F7, F8

Note: The National Technical Committee for Hydric Soils has approved the use of these field indicators for the appropriate LRR(s).

indicators, such as accumulation of muck, were good virtually nationwide but the required thickness varied by both latitude and longitude. For these reasons, field indicators have been selected for use in each LRR (Figure 9.2, Table 9.9) of the United States, Pacific Islands, and Puerto Rico. These are the only field indicators approved by NTCHS for each specific LRR. Table 9.10 provides a list of field indicators for testing by LRR.

System for Regionalizing Field Indicators

Field indicators were developed for very wet conditions nationwide by observing the centers or wettest portions of wetlands. In addition, by observing the delineation edge of ecological wetlands throughout the nation, field indicators for edges were developed region

TABLE 9.10

Test Field Indicators of Hydric Soils by Land Resource Region (LRR)

LRR	Field Indicators
A	A10
B	A10, F13
C	A9, F18 (MLRA 14)
D	F12
E	A10
F	F18 (MLRA 56)
G	S7
H	None
I	A9
J	A9, F18 (MLRA 86)
K	A10, S3, S7, S8, S9, F12
L	A10, S3, S7, S8, S9, F12
M	F12
N	None
O	A9, F18 (MLRA 131)
P	F18 (MLRA 135), F19
Q	A5
R	A10, S3, F12, F19, TA6 (MLRAs 144A and 145)
S	A10, A16 (except MLRA 149B), F19 (except MLRA 148A and 148), TA6 (MLRA 149B)
T	F19, F29 (MLRA 153B)
U	None
V	A15
W	A10, A11, F6, F7, F8, TA4, TA5
X	A10, A11, F6, F7, F8, TA4, TA5
Y	A10, A11, F6, F7, F8, TA4, TA5
Z	A5

Note: The National Technical Committee for hydric soils has approved the testing of these field indicators for the appropriate LRR(s).

by region. Other than the exceptions specified above, few of the field indicators can be used for delineation nationwide. More than 40 field indicators have been developed for use and testing; however, rarely will more than a few field indicators be used for delineation purposes in any specific region.

For example, 21 of the field indicators are identified for use in LRR N, an area that includes portions of Arkansas, Missouri, Alabama, Georgia, North and South Carolina, Tennessee, Kentucky, Virginia, West Virginia, Pennsylvania, Maryland, Ohio, and Indiana. Of these 21, seven occur predominantly in very wet areas (A1, A2, A3, A4, A12, S4, and F2) and are rarely used to delineate wetlands. Of the remaining 14, two (F6 and F7) are useful primarily for delineating wet Mollisols (soils with thick dark surface layers), two (A10 and S1) are for delineating muck or mucky soils, and three others (S5, S6, S7) are for sandy soils only. Wet Mollisols, muck, mucky, and sandy soils are rare in LRR N. Two field indicators F13 and F21 are used in one MLRA each. Therefore, to delineate most hydric soils in LRR N the most common field indicators with which one must become proficient are five (A5, A11, F3, F8, and F12).

Twenty field indicators are for use or testing in California's LRR C. Of these, seven identify very wet conditions (A1, A2, A3, A4, A12, S4, and F5). Therefore, the number of field indicators with which one must become proficient is thirteen for all of LRR C. If one's area of interest is Sacramento and San Joaquin Valleys, the number of field indicators needed is only five (F3, F6, F8, F9, and F12). As a result, it is not necessary to become familiar with all of the 40 plus different field indicators. It is only necessary to become familiar with the few field indicators used to delineate hydric soils in any given area.

Hydric Soil Determination and Delineation

Chapter 8 provides detailed information on describing soils in the field for the purpose of identifying the presence or absence of field indicators. The process of using the soil descriptions to delineate the hydric soil boundary is described in the following paragraphs.

The process of delineating hydric soil boundaries on undisturbed landscapes is really rather simple in concept but can be difficult in practice. Where the landscape is undisturbed, the upland boundary of hydric soils typically occurs at a landform change. That change is usually a convex/concave slope break. Hydric soils occur at the concave slope change, and soils that are non-hydric occur at the convex slope change. The slope break may be very subtle or hidden with vegetation, but it will be there. Often the boundary delineates a very intricate pattern of extremely small areas of hydric soils and soils that are non-hydric.

The easiest way to delineate hydric soils is to begin on the upland side of a wetland and traverse toward the wetland looking for concave slope breaks. Not all concave slope breaks delineate hydric soils; however, the hydric/non-hydric boundary of undisturbed soils will usually be at a concave slope break (see the section on Disturbed Soils for an explanation of how to delineate these soils). By traversing once or twice, the hydric soil boundary can frequently be located. Once the boundary is located, using vegetation is most expeditious for delineating the boundary over large areas and locating additional soil pits. Most often, if vegetation is present, one or two species can be correlated to the hydric soil boundary and thereby provide the key to a correct delineation. For example, in the flatwoods and associated landform areas of the southeastern United States (LRRs T and U), the uplands have the shrub saw palmetto (*Serenoa repens* L.), which disappears near the hydric soil boundary to be replaced by other shrubs, such as gallberry (*Ilex glabra* L.), and fetterbush *(Lyonia lucida* L.), in LRR T or by herbaceous plants, such as blue maidencane (*Amphicarpum mulenbergianum* L.), in LRR U.

Where vegetation is absent, the landform change from convex to concave slope break should be used to complete the delineation. Understanding that the field indicators are known to identify hydric soils is important. They were developed by observing soil pedons both inside and outside ecological wetlands. Pedons inside the line were described; descriptions of pedons outside the line were not deemed necessary. For example, S7 (Dark Surface) requires a layer 10 cm (4 in.) or more thick starting within the upper 15 cm (6 in.) of the soil surface with a matrix value 3 or less and chroma 1 or less. In this layer at least 70% of the visible soil particles must be covered, coated, or similarly masked with organic material, and the matrix color of the layer immediately below the dark layer must have chroma 2 or less. This does not mean that the pedons outside the hydric soil boundary had all requirements of this field indicator except thickness of the dark surface. It means that, because of the concave slope break, pedons outside the line are normally very dissimilar to pedons inside the line. Normally, neighboring pedons outside the line have a surface layer that has a salt-and-pepper appearance and is more of a 50/50 mixture of soil material masked with organic material (pepper) and soil material not masked (salt).

Vertical and Horizontal Soil Variability

Soil variability occurs vertically within a soil and horizontally across the landscape. Vertical variability is related to depositional and soil-forming processes. Horizontal variability is related to vertical variability and to site-specific landscape expressions of geomorphic processes; therefore, in most soils that represent simple landforms, soil variability is relatively unimportant in making a hydric soil determination. Most soils identified to be hydric on a specific landform are hydric throughout the extent of that landform, and where an upland is encountered the soils are no longer hydric. Soils have a vertical sequence of horizons that have perceptible and predictable changes with depth; however, a significant portion of soils with high shrink/swell potential are different. In the section "Problem Hydric Soil Delineations," we will discuss high shrink/swell soils and other difficult to delineate hydric soils.

Discharge Recharge and Flowthrough Hydric Soils

Discharge hydric soils release groundwater to the land surface through springs, seeps, fens, and other discharge zones including uplands (Chapter 3). Recharge hydric soils transmit water to the groundwater/aquifer and to discharge hydric soils. Hydric soils in the humid southeastern and eastern United States generally are recharge hydric soils; however, they may function as season dependent discharge systems. Both recharge and discharge hydric soils exist in the subhumid Midwest, Southwest, and West of the United States (see Chapter 3 for more discussion concerning this topic). The significance to hydric soils is that discharge systems generally have different morphological characteristics than recharge systems. Classic discharge hydric soils have morphologies that reflect water moving to the soil surface. This water carries materials, such as reduced Fe, reduced Mn, and calcium and these become part of the soil. Discharge hydric soils may lack evidence of saturation below a depth of about 0.5 m because of additions of Fe from ground water and low available organic matter needed for microbial activity. The following are examples of discharge field indicators: A1 (Histostols and Histels) in fens A16 (Coast Prairie Redox), S5 (Sandy Redox), S11 (High Chroma Sands), F3 (Depleted Matrix) where the depleted matrix is the near-surface layer, F6 (Redox Dark Surface), F8 (Redox Depressions), F9 (Vernal Pools), F16 (High Plains Depressions), and F18 (Reduced Vertic).

Recharge hydric soils are wet throughout and remain wet as long or longer than discharge hydric soils. The amount of organic matter and microbial activity is very high, and these hydric soils have maximum expressions of anaerobiosis. Recharge activities often leach soils, creating acid Fe-depleted soils. The acidity may be reflected in plants that produce tannin. Tannins in turn create organic surfaces that aid in holding water for anaerobiosis. Recharge field indicators include A1 (Histosols and Histels) in bogs, A3 (Black Histic), A12 (Thick Dark Surface), S2 (Sandy Gleyed Matrix), S6 (Stripped Matrix), S7 (Dark Surface), S8 (Polyvalue Below Surface), S9 (Thin Dark Surface), F2 (Loamy Gleyed Matrix), F3 (Depleted Matrix), where the depleted matrix is not the near-surface layer and is continuous, F7 (Depleted Dark Surface), F13 (Umbric Surface), and F22 (Very Shallow Dark Surface).

Flowthrough hydric soils transmit water to other wetter hydric soils or bodies of water. Water is transmitted by over land flow. Classic flowthrough field indicators are A5

(Stratified Layers), S6 (Organic Bodies), F10 (Marl), F12 (Iron/Manganese Masses), and F19 (Piedmont Flood Plain Soils).

Field indicators not specified as one of the discharge or recharge field indicators above have either discharge/recharge dependent morphologies or they are for hydric soils that function as season-dependent discharge and recharge hydric soils. For example, field indicator F21 (Red Parent Material) occurs in discharge, recharge, and flowthrough hydric soils. It is recommended that delineators evaluate the hydrologic source and examine soils accordingly.

Problem Hydric Soil Delineations

High Shrink/Swell Potential Soils

High shrink/swell soils are partially defined as having high (>30%) clay content, which restricts the movement of water into the soil. Most soils with high shrink/swell potential (Vertisols) (e.g., Sharkey series) have micro-variability within a soil body (pedon). Vertical sequence (horizons or layers) and horizontal variability vary greatly within a short distance from any point (Williams et al. 1996). For determining the hydric status of Vertisols, understanding this variability is important. As a result of the slickensides or wedge-shaped aggregates, Vertisols have micro-lows and micro-highs that are approximately 2–5 m from the centers of the lows to the centers of the highs. A maximum expression of the subsurface highs and lows is gilgai (U.S. Department of Agriculture 1975). Micro-lows have more organic carbon, less gypsum and carbonates, a higher coefficient of linear extensibility, and a higher probability of being hydric than micro-highs.

High shrink/swell hydric soils (Vertisols) are hydric because of surface inundation. Vertisols become hydric where water remains on the surface long enough for anaerobic bacteria to deplete the soil water of O_2. Rarely does the resulting anaerobiosis penetrate into the soil to a significant depth. Therefore, it is important to look at near-surface soil morphologies to determine the hydric status of Vertisols. Vertisols in depressions (both large scale and micro-lows) and other concave landforms are most often hydric. Vertisols in micro-highs are most often non-hydric.

The exact extent of hydric Vertisols in a particular area is highly variable. For example, hydric Vertisols of the Alabama, Mississippi, and Arkansas Blackland Prairies region of LRR P occur exclusively on depressional landforms of floodplains and lack significant gilgai relief. In these soils, the field indicator F3 (Depleted Matrix) is used almost exclusively to delineate hydric Vertisols, where the depleted matrix is within 25 cm (10 in.) of the mineral soil surface the soil is hydric. Vertisols that are non-hydric have a depleted matrix starting below 25 cm (10 in.) or have chroma 3 or more in a surface layer that is more than 15 cm (6 in.) thick. Approximately 5%–25% of the Vertisols occur in depressional positions on the floodplains of the Alabama, Mississippi, and Arkansas Blackland Prairies (LRR P) and are hydric; the remaining 75%–95% are non-hydric.

Hydric Vertisols in the Mississippi River Delta of LRR O also occur on depressional landscape positions. Field indicator F3 (Depleted Matrix) is also used to delineate these Vertisols. The depressional hydric Vertisols of the Mississippi River Delta are easy to delineate; most often they pond water for much of the year. However, Vertisols in this area that do not occur in depressions are difficult to delineate. Large areas may be either entirely hydric or entirely non-hydric, and large areas may have hydric and non-hydric soils so intricately mixed that separation of individual areas of hydric and non-hydric soils is extremely difficult. All of the Vertisols occurring in depressional landforms in the Mississippi Delta are hydric. Approximately, 70% of the Vertisols that are not in depressional landforms are hydric. Rises, knolls, and micro-highs in gilgai Vertisols normally lack field indicator F3 (Depleted Matrix).

Playas

The term *playa* is used to describe two different conditions. One condition occurs in the saline and alkaline flats of LRR D, extending from Oregon to New Mexico. Unvegetated areas that include unvegetated playas, beaches, rock outcrops, riverwash, salt flats, slickens, and slickspots may lack morphologies characteristic of hydric soils (U.S. Department of Agriculture 1993), may not have field indicators, but may be considered other waters of the United States (Environmental Laboratory 1987). The other condition occurs in the depressional landscapes of the High Plains in LRR H from Nebraska to Texas (Mitsch and Gosselink 1993). Saline playas and high plains playas (also called depressions) are hydric due to wetness from surface water and not from below-ground saturation. Therefore, the field indicators used to delineate them are based on near-surface morphological characteristics.

Saline Playas

Playas in LRR D are either sparsely vegetated or they are not currently capable of producing vegetation because of high salinity and/or alkalinity. Playas range in size from less than a hectare to many thousands of hectares. These areas are characteristically lacking in any significant pedological development or morphology that result in the development of field indicators. Although the nonvegetated areas are not considered soils by accepted USDA definition, magnetic susceptibility technology has been proven to delineate the boundary of hydric conditions in saline playas. Magnetic susceptibility (MS) is a measure of the amount of magnetic minerals (e.g., magnetite) as soil contains (Grimley and Vepraskas 2000). MS readings are expressed as 10^{-5} m^3/kg. In an unpublished study by the authors, the vegetative edge of saline playas had minimum MS values of 100–120 m^3/kg whereas values above 200 m^3/kg indicated non-hydric areas. Similar results have been reported in the upper Midwest; however, the low range of MS values in North Carolina limits MS technology there (Grimley and Vepraskas 2000) and Florida (Zwanka et al. 2007).

High Plains Playas (Depressions)

The playas (depressions) of LRR H are not true saline playas. They are vegetated depressions. Two field indicators were developed to delineate the areas of hydric soils in these high plains playas. Field indicator F8 (Redox Depressions) is restricted to use in closed depressions subject to ponding. Hydric soils are recognized where 5% or more distinct or prominent redox concentrations occur as soft masses or pore linings in a layer 5 cm (2 in.) or more thick entirely within the upper 15 cm (6 in.) of the soil surface. Field indicator F16 (High Plains Depressions) is also restricted to use in closed depressions subject to ponding. This field indicator is used to recognize hydric soils where a layer at least 10 cm (4 in.) thick within the upper 35 cm (13.5 in.) of the mineral soil has a chroma 1 or less and 1% or more redox concentrations as nodules or concretions with distinct or prominent coronas. These two field indicators (Table 9.1) along with field indicators F3 and F6 are most often used to differentiate the hydric playas from non-hydric playas. Field indicator F8 occurs most often in Texas and Oklahoma, and F16 most often in Kansas and Nebraska. F3 and F6 occur throughout the United States.

Soils with Red Parent Material

Soils with red parent material often present a delineation challenge. These soils occur in areas such as the Triassic/Jurassic sediments in the Connecticut River valley, the Permian "Red Beds" in Kansas, clayey red till and associated lacustrine deposits around the Great Lakes, Jurassic sediments associated with "hogbacks" on the eastern edge of the Rocky Mountains, and river alluvium of rivers such as the Red, Congaree, Chattahoochee, and Tennessee. Rabenhorst and Parikh (2000) developed a color change propensity index (CCPI) to help identify the presence of red parent material. The field indicator F21 (Red Parent Material) was developed specifically for hydric soil delineation in areas with red parent material in the mid-Atlantic region. Other field indicators useful in delineating the hydric component of soils that formed in red parent material include F8 (Redox Depression), F9 (Vernal Pools), F12 (Iron/Manganese Masses), F6 (Redox Dark Surface), and F3 (Depleted Matrix), where the depleted matrix is within 15 cm (6 in.) of the soil surface.

Disturbed Soils

Identifying and delineating hydric soils in areas that have been filled, dredged, land leveled, or otherwise disturbed can be difficult and extremely challenging. In some instances of disturbance, the vegetation has been destroyed or removed; therefore, soils are the only onsite field indicator of predisturbance hydrology and the only feasible means of identifying wetlands. Where upturned soil disturbance is recent, sufficient clods of various soil horizons may remain that will aid experienced soil scientists in verifying the original soil morphology. Predisturbance soil surveys should be consulted where available. Undisturbed areas in the vicinity may be investigated to provide information of predisturbance soil morphology. Small areas of unaltered soil may be found at the base of remaining trees; however, most frequently, the disturbance is more extreme. Fill materials spread on disturbed sites usually compound the difficulties of making hydric soil determinations. An experienced soil scientist can often identify the contact between fill material and the original soil surface. Guidelines have been established to determine the hydric status of disturbed soils after varying amounts of fill materials have been added. These guidelines are based on insights and observations of the authors and are not to be considered official guidance for CWA and FSA use.

Hydric soil requirements are the same for disturbed areas as they are for undisturbed areas. Most significantly, the hydric soil definition must be satisfied (*Federal Register,* July 13, 1994). This is normally exemplified by the presence of a field indicator (Hurt and Carlisle 1997). The amount of fill that can be placed on a hydric soil and still allow that soil to be considered hydric is directly related to the field indicator and the reason (inundation or saturation) it was hydric prior to filling.

For areas that meet the requirements for Histosols (except Folists, Histels, and Folistels) fill can be placed on the soil surface to the depth that the soil, after the placement of the fill, still meets the taxonomic requirements of Histosols (U.S. Department of Agriculture 1996). Therefore, the maximum amount of fill material that can be added to a hydric Histosol and still have that soil retain its hydric status is 40 cm (16 in.) (60 cm if 3/4 or more of the organic soil material is moss fibers). This would apply to hydric Histosols that have organic soil material starting at the soil surface that is 40 cm (16 in.) or more thick (60 cm or more thick if 3/4 or more of the organic soil material is moss fibers). For Histosols with thinner organic layers (e.g., organic soils over bedrock), the thickness of the fill material would be less to maintain their hydric status.

For soils that are frequently ponded for long or very long duration during the growing season or soils that are frequently flooded for long or very long duration during the growing season to maintain their hydric status after filling, the thickness of the fill must be slightly less than the height of frequent ponding or flooding of long duration (more than 7 days). This height may be either measured or estimated. If estimated, professional judgment that the definition (anaerobiosis) is met must be carefully exercised. Although any of the field indicators listed in Table 9.1 may occur on inundated landforms, field indicators A4 (Hydrogen Sulfide), F8 (Redox Depressions), F9 (Vernal Pools), F12 (Iron/Manganese Masses), F13 (Umbric Surface), and F16 (High Plains Depressions) are restricted to landforms subject to inundation.

For soils that are hydric due to saturation (USDA NRCS 2010) the depth of fill that can be placed on these soils in order to maintain their hydric status is variable. The range is from slightly less than 15 cm (6 in.) to 0 cm in soils with sandy soil materials and the range is from slightly less than 30 cm (12 in.) to 0 cm in other soils. After fill materials are added, a field indicator must be present in the original soil material within the prescribed depths in order for that soil to retain its hydric status. Table 9.11 can be used to determine the depth of fill material that would adversely affect the hydric status of a soil that is hydric due to saturation. This table is not to be used for field indicator A1 and the field indicators restricted to landforms subject to inundation (e.g., F8, F9, F11, F12, and F16).

Soils with a field indicator starting depth that is intermediate to those listed in Table 9.11 can have an intermediate amount of fill without changing the hydric status of the soil. For example, a soil with a stripped matrix starting at 10 cm (4 in.) can have up to 5 cm (2 in.)

TABLE 9.11

Hydric Status of Soils with Varying Amounts of Fill[a]

Original Depth to Field Indicator	Type of Field Indicator[b]	Thickness of Fill Material	Hydric Status
Surface	All, Sandy	Up to 15 cm (6 in.)	Hydric
Surface	All, Sandy	More than 15 cm (6 in.)	Non-hydric
Surface	Loamy or Clayey	Up to 30 cm[c] (12 in.)	Hydric
Surface	Loamy or Clayey	More than 30 cm[c] (12 in.)	Non-hydric
15 cm	All, Sandy	Zero	Hydric
15 cm[d]	All, Sandy	More than zero	Non-hydric
15 cm	Loamy or Clayey	Up to 15 cm[d] (6 in.)	Hydric
30 cm[c]	Loamy or Clayey	Zero	Hydric
30 cm[c]	Loamy of Clayey	More than zero	Non-hydric

[a] This table is used to determine the depth of fill material that would adversely affect the hydric status of a soil that is hydric due to saturation and is based on the presence or absence of a field indicator; however, if a field indicator is absent, a soil may well be hydric if, according to NTCHS guidance, the definition is met.

[b] See *Field Indicators of Hydric Soils in the United States* for additional information concerning use of All (A), Sandy (S), and Loamy and Clayey (F) field indicators.

[c] Depths and thicknesses would be 25 cm (10 in.) if the field indicator present is F3 (Depleted Matrix).

[d] Depth would be 10 cm (4 in.) if the field indicator present is F3 (Depleted Matrix).

of any type of fill material placed on the surface without changing the hydric status of the soil. Conversely, more than 5 cm (2 in.) of fill would change the status of the soil to non-hydric.

The procedure described for filled areas should be used to determine the hydric status of land-leveled areas. Soils that are hydric due to inundation prior to land leveling are evaluated after land leveling to determine their hydric status. Soils that are hydric due to saturation prior to land leveling are evaluated by applying the guidelines outlined in Table 9.11 to determine their hydric status.

The presence of structures that provide increased drainage (ditches, tile drains, etc.) and protection from ponding and/or flooding (dikes, levees, etc.) does not alter the hydric status of a soil.

Summary

Field indicators provide a proof positive approach that the definition of a hydric soil has been met. Field indicators are an integral part of a three factor approach of using a dominance of hydrophytic vegetation, indicators of wetland hydrology and field indicators of hydric soils to provide a practical, rapid, repeatable, science-based method to identify and delineate wetlands. Some users may find the number of field indicators to be challenging; but through a system of regionalization the number of field indicators that are likely to be found at a wetland boundary is quite manageable. Some field indicators are nearly universal in their distribution across the country (e.g., A1 Histosol) others are limited to a small sub-region (e.g., F17 Delta Ochric). In many altered settings, field indicators are the only remaining clue that a site was a wetland. The persistent nature of field indicators when vegetation and hydrology have been completely altered from a natural setting can be a useful tool in wetland restoration, by showing where the wetland boundary was before alteration. The current list of field indicators is dynamic through a published and tested process that is likely to change as new knowledge and information becomes available.

References

Cowardin, L. M., V. Carter, F. C. Golet, and E. T. LaRoe. 1979. *Classification of Wetlands and Deepwater Habitats of the United States*. U.S. Fish and Wildlife Service, Washington, DC, FWS/OBS-79/31.

Environmental Laboratory. 1987. *Corps of Engineers Wetland Delineation Manual.* Technical Report Y-87-1. U.S. Army Engineers Waterways Experiment Station, Vicksburg, MS.

Faulkner, S. P., W. H. Patrick, Jr., R. P. Gambrell, W. B. Parker, and B. J. Good. 1991. Characterization of soil processes in bottomland hardwood wetland–nonwetland transition zones in the lower Mississippi River Valley. U.S. Army Corps of Engineers, Waterways Experiment Station, WRP-91-1. Vicksburg, MS.

Federal Register. July 13, 1994. *Changes in Hydric Soils of the United States*. National Archives and Records Administration and U.S. Government Printing Office, Washington, DC.

Federal Register. February 29, 2012. *Changes in Hydric Soils Database Selection Criteria*. National Archives and Records Administration and U.S. Government Printing Office, Washington, DC.

Florida Soil Survey Staff. 1992. *Soil and Water Relationships of Florida's Ecological Communities*. G. W. Hurt (Ed.). USDA, Soil Conservation Service, Gainesville, FL, 16 pages.

Grimley, D. A. and M. J. Vepraskas, 2000. Utilization of magnetic susceptibility for delineating hydric soils. *Soil Science Society of America Journal* 64: 2174–2180.

Hammer, D. A. 1992. *Creating Freshwater Wetlands*. Lewis Publishers, Boca Raton, FL, pp. 20–37.

http://soils.usda.gov/use/hydric/Hydric soil technical standard, technical note 4.

http://soils.usda.gov/use/hydric/Hydric soil technical standard, technical note 11.

Hurt, G. W. and R. B. Brown. 1995. Development and application of hydric soil indicators in Florida. *Wetlands* 15(1): 74–81.

Hurt, G. W. and V. W. Carlisle. 1997. Proper use of hydric soil terminology. *Soil Survey Horizons* 38(4): 98–101.

Lichvar, R. W. and J. T. Kartesz. 2009. North American digital flora, national wetland plant list version 24.0 (http://wetland_plants.usace.army.mil). U.S. Army Corps of Engineering, Engineer Research and Development Center. Cold Regions Research and Engineering Laboratory, Hanover, NH, and BONAP, Chapel Hill, NC.

Lichvar, R. W. and P. Munkin. 2008. *Concepts and procedures for updating the national wetland plant list*. No. ERDC/CRREL-TN-08-3. Cold Regions Research and Engineering Laboratory, Hanover, NH.

Mausbach, M. J. and J. L. Richardson. 1994. Biogeochemical processes in hydric soils. In *Wetland Biogeochemistry, Volume 1.* Wetlands Biogeochemistry Institute, Louisiana State University, Baton Rouge, LA, pp. 68–127.

Mitsch, W. J. and J. G. Gosselink. 1993. *Wetlands.* Van Nostrand Reinhold, New York.

Rabenhorst, M. C. and S. Parikh. 2000. Propensity of soils to develop redoximorphic color changes. *Soil Sci. Soc. Am. J.* 64(5): 1904–1910.

Segal, S. D., S. W. Sprecher, and F. C. Watts. 1995. Relationships between hydric soil indicators and wetland hydrology for sand soils in Florida. U.S. Army Corps of Engineers, Waterways Experiment Station, WRP-DE-7. Vicksburg, MS.

Skaggs, R. W., D. Amatya, R. O. Evans, and J. E. Parsons. 1991. Methods of evaluating wetland hydrology. ASAE Paper No. 912590, American Society of Agricultural Engineering, St. Joseph, MI.

Skaggs, R. W., D. Amatya, R. O. Evans, and J. E. Parsons. 1994. Characterization and evaluation of proposed hydrology criteria for wetlands. *J. Soil Water Conserv.* 49(5): 501–510.

Sumner, J. P., M. J. Vepraskas, and R. J. Kolka. 2009. Methods to evaluate rainfall for short-term hydrology assessment. *Wetlands* 29(3): 1049–1062.

Tiner, R. W. and D. G. Burke. 1995. *Wetlands of Maryland*. U.S. Fish and Wildlife Service, Hadley, MA and Maryland Department of Natural Resources, Annapolis, MD, Cooperative publication, p. 7.

Tuttle, S. D. 1975. *Landforms and Landscapes.* Wm. C. Brown Company, Dubuque, IA.

USDA NRCS. 2010. Field indicators of hydric soils in the United States, Version 7.0. In L. M. Vasilas, G. W. Hurt, and C. V. Noble (Eds.) *USDA, NRCS, in cooperation with the National Technical Committee for Hydric Soils*.

U.S. Department of Agriculture, Soil Conservation Service. 1981. *Land Resource Regions and Major Land Resource Areas of the United States.* USDA–SCS Agricultural Handbook 296. U.S. Govt. Printing Office, Washington, DC.

U.S. Department of Agriculture, Soil Conservation Service. 1994. *National Food Security Act Manual.* 3rd edition. USDA, NRCS, Washington, DC.

U.S. Department of Agriculture, Soil Survey Staff. 1993. *National Soil Survey Handbook.* USDA, Soil Conservation Service, U.S. Govt. Printing Office, Washington, DC.

U.S. Department of Agriculture, Soil Survey Staff. 1975. *Soil Taxonomy: A Basic System of Soil Classification for Making and Interpreting Soil Surveys.* USDA Agricultural Handbook 436. U.S. Govt. Printing Office, Washington, DC.

U.S. Department of Agriculture, Soil Survey Staff. 1996. *Keys to Soil Taxonomy,* 7th edition. U.S. Govt. Printing Office, Washington, DC.

U.S. Department of Agriculture, Soil Survey Staff. 1999. *Soil taxonomy: A basic system of soil classification for making and interpreting soil surveys*. 2nd edition. Natural Resources Conservation Service. U.S. Department of Agriculture Handbook 436, Washington, DC.

Vepraskas, M.J. 1994. *Redoximorphic Features for Identifying Aquic Conditions.* Tech. Bulletin 301. North Carolina Agricultural Research Service, North Carolina State University, Raleigh, NC.

Williams, D., T. Cook, W. Lynn, and H. Eswaran. 1996. Estimating the field morphology of Vertisols. *Soil Survey Horizons* 37(4): 123–131.

Xrite. 2009. *Munsell® Soil Color Charts: Revised Edition*. Xrite, Grand Rapids, MI.

Zwanka, W. P., G. W. Hurt, D. A. Gratz, W. D. Graham, and M. W. Clark. 2007. Using magnetic susceptibility to delineate hydric soils in southeastern coastal plain soils. *Soil Survey Horizons* 48(2): 36–40.

Section II

Wetland Soil Landscapes

10

Soils of Peatlands: Histosols and Gelisols

Randy Kolka, Scott D. Bridgham, and Chien-Lu Ping

CONTENTS

Introduction

Peatlands are a subset of wetlands that have accumulated significant amounts of soil organic matter. Soils of peatlands are colloquially known as peat, with mucks referring to peats that are decomposed to the point that the original plant remains are altered beyond recognition (Chapter 6, SSSA 2008). Generally, soils with a surface organic layer >40 cm thick have been classified as Histosols in the U.S. soil classification system—Soil Taxonomy (Soil Survey Staff 2014). Permafrost-affected organic soils are classified as the Histels suborder in the Gelisols order (Soil Survey Staff 2014). Based on current calculations of earth's land surface of 148,940,000 km^2 and our estimate of peatland area (or the combined area of Histosols and Histels) (Table 10.1), peatlands occupy about 2.7% of the earth's surface.

Peatlands have historically been classified based on a number of criteria, such as topography, ontogeny (i.e., landscape developmental sequence), hydrology, soil and/or water chemistry, plant community composition, and degree of soil organic matter decomposition (Moore and Bellamy 1974; Cowardin et al. 1979; Gore 1983; Bridgham et al. 1996; National Wetlands Working Group 1997; Inisheva 2006; Vitt 2006). Given the confusion in peatland terminology and the emphasis of this chapter on soils, we will discuss here only the dominant ecological paradigm in peatlands—the ombrogenous–minerogenous gradient. Although the fundamental definition of this gradient is based on hydrology, it is often

TABLE 10.1

Current and Historical Global Peatland Area (in 10^3 km^2)

Regions[a]	Current	Historical
Alaska[b]	132	132
Canada[b]	1136	1150
Mexico[b]	10	–
US[c,b]	93	111
North America[b]	1372	1407
Northern	3728[d]	4045[e]
Tropical	285[f]	441[g]
Global	4013	4486

[a] Includes both permafrost and non-permafrost peatlands.
[b] Bridgham et al. (2006).
[c] Not including Alaska.
[d] Historical area—loss of 316,000 km^2 reported in Joosten (2009).
[e] Includes all non-tropical peatlands in Northern and Southern Hemispheres (Yu et al. 2010).
[f] Historical area—loss of 156,000 km^2 reported in Joosten (2009).
[g] Page et al. (2011).

thought to be coincident with (and a primary control over) plant community composition and the biogeochemistry of peatland soils (Bridgham et al. 1996).

Minerogenous peatlands have significant inputs of groundwater and/or upland run-off, generally imparting higher basic cation content and pH to their soils (Heinselman 1963; Moore and Bellamy 1974). These peatlands are generally called fens, whereas treed minerogenous peatlands are often termed swamp forests in North America, although this latter term is also used to describe forested wetlands on mineral soils (National Wetlands Working Group 1997). In contrast, ombrogenous peatlands, through deep accumulation of peat, have achieved a landscape topographic position where they are isolated from all but atmospheric inputs of water, alkalinity-generating cations, and nutrients. As a result, they have low ash and basic cation content and low pH in their soils, and are commonly termed bogs. Fens exhibit a wide range of minerotrophy due to complicated interactions between hydrology, topographic landscape position, and chemistry of surrounding and/or underlying mineral soils and groundwater (Bridgham and Richardson 1993; Bridgham et al. 1996; Verry 1997, 2006). For example, a region where mineral soils are dominated by sand with very low exchangeable cations can have fens with significant groundwater input but soil chemistry and plant communities more characteristic of bogs.

Fens with more minerogenous characteristics (i.e., higher soil pH and basic cation content) are generally described as "rich," whereas those more similar to bogs in soil chemistry and plant community composition are called "poor." Bridgham et al. (1996) objected to terms such as rich and poor fens, because they essentially describe a gradient of pH and basic cation concentration, while most studies have pointed to nitrogen and/or phosphorus as the limiting nutrients for plant growth in peatlands. They suggested that nutrient availability gradients may not be coincident with the ombrogenous–minerogenous gradient; experimental results have demonstrated that nitrogen availability is greater in more minerogenous peatlands, whereas phosphorus availability is higher in more ombrogenous peatlands (Bridgham et al. 1998; Chapin 1998), although recent research using enzymes as indicators of nutrient availability indicated that phosphorus was more limited than nitrogen across a ombrogenous–minerogenous gradient in northern Minnesota (Hill et al. 2014).

The effect of permafrost on peatlands is dramatic, lending support to defining the soil suborder Histels for permafrost-affected organic soils. The formation and development of several major peatland types are the direct result of permafrost action (Zoltai and Tarnocai 1971; Moore and Bellamy 1974; National Wetlands Working Group 1988; Botch et al. 1995; Ahrens et al. 2004). Additionally, soil carbon pool sizes, distribution, and bioavailability are strongly affected by (1) cryoturbation, which is the soil-mixing action of freeze/thaw processes, and (2) by the presence of permafrost itself, which has strong controls over soil temperature and moisture and runoff (Michaelson et al. 1996). Overall, permafrost-affected soils represent 16% of all soils on the globe, and contain up to 50% of the global belowground soil carbon pool (Tarnocai et al. 2009).

The literature on peatlands is vast, and we focus here only on the soils, particularly within the context of the ombrogenous–minerogenous gradient and the effects of permafrost. The objectives of this chapter are to: (1) summarize the geographic distribution of the world's peatlands, (2) describe Gelisols as defined in Soil Taxonomy and compare it to classifications of other countries and organizations, (3) examine the effects of the physical structure and botanical composition of various peats on their hydrologic properties, and (4) compare the physical and chemical characteristics of peats in U.S. wetlands from Florida to Alaska, with an emphasis on the ombrogenous–minerogenous gradient for Histosols and the defining characteristics due to permafrost in Gelisols.

Geographic Distribution

Global Peatlands

Ground-based estimates, remote sensing, and hydrological modeling have all been used to estimate the regional distribution and global area of wetlands (reviewed in Lehner and Döll 2004; Bridgham et al. 2006; Melton et al. 2013), but it is likely that ground-based estimates most effectively delineate peatlands from other wetland types (cf. Lehner and Döll 2004). There are two distinct peaks of wetland area in the tropical and boreal zones, with tropical wetlands being primarily mineral-soil based and boreal wetlands being primarily peatlands (Figure 10.1). It is interesting that, while northern climates are clearly conducive to peat formation, large areas of tropical peatlands do exist (Table 10.1)—for example, very deep peat deposits occur in Indonesia and the Amazon (Page et al. 2011). Remote sensing techniques suggest that the area of tropical wetlands may have been formerly underestimated (Gumbricht 2012).

Table 10.2 gives the distribution of organic soils within the U.S. There are two related databases maintained by the USDA that provide the best available estimates of organic soil area in the U.S. (Soil Survey Staff 1998). MUIR (Map Unit Interpretation Record) contains digitized soil maps at a scale of 1:12,000–1:31,680, but large areas of certain states have not had soil surveys completed. This includes states such as Michigan and Minnesota that have large expanses of organic soils. In the STATSGO (State Soil Geographic) data base, other sources of information are used to estimate soil information in unmapped areas, but the scale is at 1:250,000, except Alaska, which is at 1:1,000,000. In states that are poorly mapped, STATSGO data are necessary to obtain realistic estimates of peatland area. However, because of the coarse scale, STATSGO fails to recognize many small pockets of organic soils. Consequently, it was deemed most accurate to take the highest estimate of

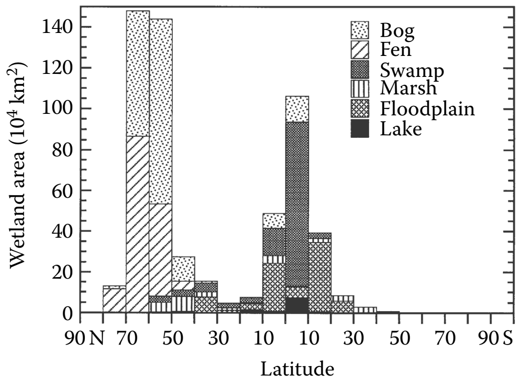


FIGURE 10.1
Global wetland area in 10° latitudinal belts for various wetland types. (Modified from Aselmann, I. and P. J. Crutzen. 1989. *J. Atmos. Chem.* 8: 307–359. With permission from Kluwer Academic Publishers.)

STATSGO or MUIR for each state (Soil Survey Staff 1998). Total organic soil area in the U.S. is 234,006 km^2 (Table 10.2), with Alaska alone accounting for 56% of all peatlands. Excluding Alaska, the two regions with the most organic soils are the Midwest and South. In particular, large areas of peatlands occur in Michigan, Minnesota, Wisconsin, Florida, Louisiana, and North Carolina.

The distribution of Alaskan peatlands into Histosols and Gelisols demonstrates that 67% of its peatlands are affected by permafrost. Tarnocai (1998) estimated that 36% of Canadian peatlands had permafrost features (Organic Cryosols). In particular, significant areas of peatlands occur in the zone of discontinuous permafrost (Gorham 1991). Mosses and black spruce tend to enhance permafrost formation in this discontinuous zone (Van Cleve et al. 1991; Camill and Clark 1998).

Global Carbon Storage in Peatlands

Although peatlands only occupy approximately 2.7% of the terrestrial land surface, they represent a globally significant carbon pool because of the deep organic soil deposits that have accumulated over thousands of years. Gorham (1991) estimated that boreal and subarctic peatlands contain 455 Pg C (1 Pg = 10^{15} g). This is very similar to the global peatland carbon pool of 462 Pg estimated by Bridgham et al. (2006). In comparison, Yu et al. (2010) estimated that Northern Hemisphere boreal and subarctic peatlands contain 547 Pg C, tropical peatlands contain 50 Pg C, and Patagonia peatlands contain 15 Pg C, for a total of 612 Pg C. Histels alone are estimated to contain 184 Pg C (Tarnocai et al. 2009), and thus contain a substantial fraction of world's peatland carbon.

Peat deposits of the boreal region tend to be deeper than those of the subarctic, and the boreal region has higher long-term net carbon accumulation rates (Ovenden 1990; Gorham 1991; Botch et al. 1995; Ping et al. 1997a; Bridgham et al. 2006; Kolka et al. 2011). On average, long-term accumulation rates in subarctic and boreal peatlands were estimated to be 7–11 and 23–41 g m^{-2} yr^{-1}, respectively (Ovenden 1990). Carbon accumulation rates ranged from 12 g m^{-2} yr^{-1} in Arctic peatlands to 80 g m^{-2} yr^{-1} in more minerotrophic mires in the boreal and temperate zones of the former Soviet Union, with an average of 30 g m^{-2} yr^{-1} (Botch et al. 1995). Bridgham et al. (2006) estimated the mean carbon accumulation rate to

TABLE 10.2

Area of Organic Soils (km^2) in the United States

State	Histosol	Data[a]	Histel	State	Histosol	Data	Histel
Midwest				**South**			
Illinois	356	M	–	Alabama	809	S	–
Indiana	1490	S	–	Arkansas	–	M	–
Iowa	301	M	–	Florida	15,943	S	–
Kansas	–	M	–	Georgia	1879	S	–
Michigan	16,511	S	–	Kentucky	–	M	–
Minnesota	24,345	S	–	Louisiana	9537	M	–
Missouri	51	M	–	Mississippi	908	S	–
Nebraska	44	M	–	North Carolina	6339	S	–
North Dakota	26	M	–	Puerto Rico	28	M	–
Oklahoma	–	M	–	South Carolina	650	S	–
South Dakota	–	M	–	Tennessee	–	M	–
Wisconsin	13,476	S	–	Texas	52	M	–
Total	56,601			Virginia	549	S	–
				Total	36,693		
Northeast				**West**			
Connecticut	434	S	–	Alaska	43,201	S	88,994
Delaware	356	S	–	Arizona	–	M	–
Maine	3965	S	–	California	617	S	–
Maryland	949	M	–	Colorado	335	S	–
Massachusetts	1364	M	–	Hawaii	1920	M	–
New Hampshire	899	M	–	Idaho	236	S	–
New Jersey	732	M	–	Montana	260	S	–
New York	3131	S	–	Nevada	74	S	–
Ohio	309	S	–	New Mexico	1	M	–
Pennsylvania	163	M	–	Oregon	329	S	–
Rhode Island	119	S	–	Utah	28	S	–
Vermont	270	M	–	Washington	790	M	–
West Virginia	–	M	–	Wyoming	30	M	–
Total	12,692			Total	47,821		88,994

Source: Adapted from Soil Survey Staff. 1998. *Query for Histosol Soil Components in the National MUIR and STATSGO Data Sets 8/98*. Natural Resource Conservation Service, USDA, Lincoln, NE and Statistical Laboratory, Iowa State University, Ames, IA.

Note:
Total Peatlands = 234,006.
Total Histosols = 153,807.
Total Wetland Histosols = 145,012.
Total Folists = 8795 km^2.
[a] S = STATSGO, M = MUIR. The highest Histosol area was taken from either STATSGO (State Soil Geographic database) or MUIR (Map Unit Interpretation Record database). Folist and Histel area were taken from STATSGO.

be 7.1 g m^{-2} yr^{-1} for the conterminous U.S. while Kolka et al. (2011) synthesized the literature and reported a range from 0.7 to 42 g m^{-2} yr^{-1} across all Histosols with rates generally increasing with decreasing latitude.

Although peatlands are generally sinks for atmospheric carbon, they are also important sources of greenhouse gases. Wetlands are an important land use that is tracked by countries for Intergovernmental Panel on Climate Change (IPCC) reporting (IPCC 2006). Recently a Wetlands Supplement was produced by the IPCC to better account for greenhouse gas fluxes and changes in carbon pools for managed peatlands (IPCC 2014). Kolka et al. (2011) completed a synthesis of the literature for carbon dioxide and methane fluxes from peatlands across the globe. For natural or unmanaged peatlands, the mean flux of carbon dioxide was 79.5 mmol m^{-2} d^{-1} (range 12–152 mmol m^{-2} d^{-1}), while for methane it was 5.4 mmol m^{-2} d^{-1} (range 0.03–18 mmol m^{-2} d^{-1}). A number of peatland drainage experiments were also included in the synthesis and drainage tends to increase carbon dioxide fluxes by about a factor of three while decreasing methane fluxes by about a factor of three (Moore and Knowles 1989; Nykanen et al. 1995; Strack et al. 2004; Kolka et al. 2011).

Gelisols

Histosol soil classification was discussed in Chapter 6. In this section we will briefly discuss the classification of organic soils in three widely used soil classification systems. In Soil Taxonomy, the U.S. soil classification, organic soils not affected by permafrost are placed in the Histosol order and those affected by permafrost are keyed out in the Histels suborder under the Gelisol order. Great groups of Histels are defined by fiber contents, period of saturation (differentiation of histic vs. folic) and presence of ground ice. In the Canadian system (Soil Classification Working Group 1998), organic soils are recognized at the Organic order, and the ones affected by permafrost are keyed out in the Organic Cryosol great group of the Cryosolic order. Subgroups then are defined by fiber content of the control section or by the depth of peat over mineral soil or ice. In the World Reference Base system (IUSS Working Group WRB 2006), organic soils are recognized at the Reference Soil Group (RSG) as Histosol and those affected by permafrost are placed at the second level with a qualifier as Cryic Histosol. In all three systems, the requirements for Histosols and permafrost-affected Histosols (Histels) are comparable.

Comparison of Four Classification Schemes

By way of comparison, we examine four alternative methods for classifying organic soils from Florida to Minnesota and two histic epipedons from a beaver meadow (Table 10.4 and see *Peat Biogeochemistry—A Comparative Approach* below). The first method is the USDA protocol (Soil Survey Staff 2014), as described in Chapter 6. The second is the ASTM protocol (ASTM 2013), with sapric, hemic, and fibric peats having 0%–32%, 33%–67%, and >67% dry-mass unrubbed fiber, respectively. The third method is the Canadian protocol (Soil Classification Working Group 1998), with sapric peat having a rubbed fiber content of <10% by volume and a pyrophosphate index (determined on the Munsell color chart after inserting white chromatographic paper into a paste composed of peat and a sodium-pyrophosphate solution) of ≤3, fibric peat having ≥75% rubbed fiber content by volume *or* ≥40% rubbed fiber by volume and a pyrophosphate index of ≥5, and hemic peat failing to meet the requirements of fibric or sapric peat. The fourth method is the von Post scale (Mathur and Farnham 1985; Parent and Caron 1993; ASTM 2013), where sapric, hemic, and fibric peats have von Post ratings of 7–10, 4–6, and 1–3, respectively.

None of the samples had ≥75% average rubbed fiber content by volume, but most of the bog and acidic fen soils would be classified as fibric in the USDA and Canadian systems based on their pyrophosphate color. Visually these samples were composed predominantly of moderately to undecomposed *Sphagnum* fibers. Similar results were obtained with the ASTM classification system. The von Post scale gave a greater variety of classification values for bogs and acidic fens.

The intermediate fens, tamarack swamps, and cedar swamps had hemic peat according to most of the classification systems, whereas the histic epipedon in the beaver meadows, the ash swamp, and the southern peats had sapric material according to one or more of the classification systems. Correlations between the classification systems ranged from an r^2 of 0.54 (between von Post and ASTM) and 0.88 (between Canadian and ASTM). Thus, quite different classifications can be given by the different systems, even though peats are only divided into three decompositional categories. Overall, the Canadian system tended to give highest values (i.e., the fewest Saprists and Hemists), and the USDA and von Post systems the lowest values (Table 10.4).

There are 279 Histosol soil series in the U.S. (excluding Folists) (Kolka et al. 2011). Of those series, 9.3% are Fibrists, 29.4% Hemists, and 61.3% Saprists. In comparison, Canadian Histosols (their Organic order) are 36.8% Fibrists (their Fibrisol), 61.8% Hemists (their Mesisol), and only 1.4% Saprists (their Humisol; Tarnocai 1998). The differences between the two countries probably reflect greater decomposition of peats at lower latitudes (see *Peat Biogeochemistry—A Comparative Approach* below), and the tendency of the Canadian soil classification system to place similar peats into less decomposed categories than the U.S. system, as discussed above.

Malterer et al. (1992) reviewed methods of assessing fiber content and decomposition in northern peats. They compared the von Post method, the centrifugation method of the former Soviet Union (Parent and Caron 1993), the USDA pyrophosphate color test and fiber-volume methods, and the ASTM fiber-weight method. Their analyses indicate that the centrifugation method of the former Soviet Union and the von Post humification field method separate more classes of peat with greater precision than the USDA and the ASTM methods. Stanek and Silc (1977) similarly found the von Post method differentiated more classes of well-humified peat than the rubbed and unrubbed fiber volume methods and the pyrophosphate color test of the USDA.

The pyrophosphate method is not particularly effective at extracting peat humic substances (Mathur and Farnham 1985). Additionally, the use of pyrophosphate color is limited because it is a qualitative variable, although spectrophotometric alternatives exist (Day et al. 1979). Mathur and Farnham (1985) state, "There is little theoretical basis for assuming that the color intensity of a [pyrophosphate] peat extract should be closely related to the extent of humification or that the extraction would be even semiquantitative in the presence of significant amounts of mineral matter." However, the pyrophosphate color index is reasonably well correlated with other measures of humification in Table 10.5.

Hydrology

Hydrology and Peatland Development

Hydrology is the central factor, by definition, in the formation of all hydric soils, but peatlands are unique in the degree of autogenic (i.e., biotically driven) feedbacks between plant production and community composition, microbial decomposition, soil biogeochemistry, and hydrology (Heinselman 1963, 1970; Moore and Bellamy 1974; Siegel 1992; Belyea and Baird 2006). Under waterlogged conditions, especially in northern latitudes as noted above, net primary production generally exceeds decomposition, resulting in peat formation. The peat's botanical source, state of decomposition, bulk density, and depth interact to determine its hydraulic conductivity (Boelter 1969; Päivänen 1973; Silins and Rothwell 1998; Weiss et al. 1998). At some point, accumulation of deep, highly decomposed peat may impede vertical groundwater exchange with the surface layers. Additionally, the formation of peat itself increases water retention. As water retention increases, the peatland expands above the regional water table, and often above the surrounding landscape. At this point, an ombrogenous system has developed, with its characteristic soil chemistry and plant communities. Thus, we see a succession over time in many peatlands from fens to bogs, with an increasing state of ombrotrophy as a result of increasing biotic control over hydrology.

There are climatic limitations on this process: fens can occur in any climate because of their dependence on outside sources of water, whereas bogs can only occur in regions where precipitation exceeds evapotranspiration. The preponderance of peatlands in northern latitudes is at least partially due to lower temperatures limiting evapotranspiration, so that peatland formation is favored in areas of even moderate precipitation. However, substrate permeability, artesian pressure heads, landform, and other groundwater factors can override macroclimate in the formation of large peatland complexes (Heinselman 1970; Siegel and Glaser 1987). In permafrost regions, drainage is further slowed by the seasonal freeze–thaw cycle, underlying permafrost, and low evapotranspiration rates, especially on north-facing slopes (Rieger 1983) where "hanging bogs" were described. Permafrost may also act as a confining aquatard, creating artesian conditions for groundwater discharge and spring-fed wetlands (Racine and Walters 1994).

Hydrology and Peat Characteristics

As noted above, an important attribute of peats is their ability to hold and retain water. Undecomposed fibric peats are predominantly composed of air- or water-filled pore spaces of large diameter (>600 μm, Boelter 1964; Päivänen 1973; Silins and Rothwell 1998). This, in combination with low-density organic matter, results in a saturated water content often exceeding 1000% of oven-dry mass and 90% of total peat volume (Boelter 1964, 1969; Päivänen 1973; Damman and French 1987) (see Figure 10.2). More decomposed, higher bulk density peats and herbaceous peats have smaller pore spaces and correspondingly lower water-storage capacity under saturated conditions, although they still maintain >80% saturated water content by volume (Boelter 1964, 1969; Päivänen 1973; Silins and Rothwell 1998) (see Figure 10.2). However, water is held in the large pore spaces of fibric peat primarily by detention storage (i.e., easily drainable porosity), and even moderate soil tensions result in large losses of the stored water (Figure 10.2). Similar to mineral soils, more decomposed, higher bulk density peats, with correspondingly smaller diameter pore spaces, have greater water retention under unsaturated conditions, and this difference increases at higher soil tension (Figure 10.2). The different botanical compositions of peats also have an important effect on water-holding capacity and retention (Boelter 1968; Weiss et al. 1998).

Surface peats have horizontal conductivities that are orders of magnitude greater than downward hydraulic conductivities in deeper peats (Päivänen 1973; Ingram 1982, 1983; Gafni and Brooks 1990). An important cause of this anisotropy is that deeper, more decomposed peat layers tend to have lower saturated hydraulic conductivity (Figure 10.3). In peatland terminology, water flow occurs predominantly in the upper, seasonally aerobic layer of the peat, or acrotelm, with very low flow through the deeper, permanently anaerobic layer,

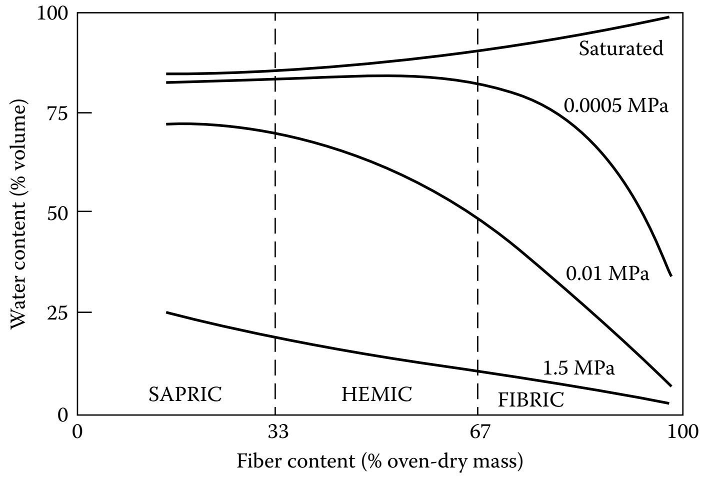


FIGURE 10.2
Relationship between fiber content of peat, water content, and soil water potential. Note that the definition for sapric, hemic, and fibric peats is somewhat different than used today in the U.S. (Modified from Boelter, D. H. 1969. *Soil Sci. Soc. Am. Proc.* 33: 606–609. With permission.)

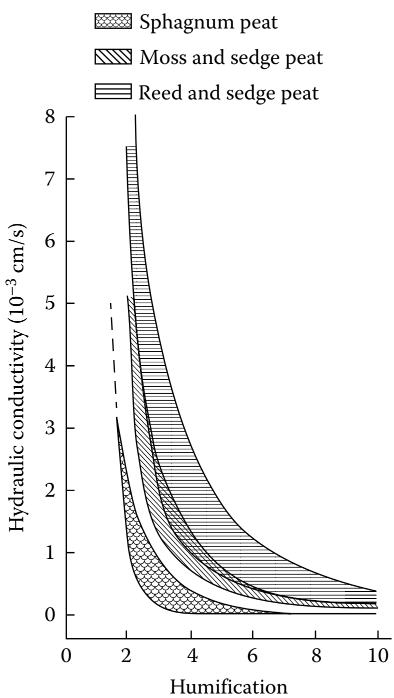


FIGURE 10.3
Effect of the botanical composition of peat and degree of humification on saturated hydraulic conductivity of peats. Humification is given in the qualitative von Post scale, where 1 is undecomposed and 10 is extremely decomposed. (Modified from Baden, W. and R. Eggelsmann. 1963. *Zulturtech. Flurber.* 4: 226–254.)

or catotelm (Damman 1986). Interestingly, unsaturated hydraulic conductivity is greater in more decomposed peats with smaller-diameter pore spaces (Silins and Rothwell 1998), similar to mineral soils (Brady and Weil 2008). Additionally, the plant composition from which the peat was derived has a dramatic effect on saturated hydraulic conductivity, with reed–sedge peat having the highest conductivity, and *Sphagnum* peats the lowest within any particular humification class (Figure 10.3). Undecayed *Sphagnum* moss has very high saturated conductivities, but conductivity decreases rapidly upon humification. Despite the high surface saturated conductivity of peats, horizontal water movement is very slow due to the low slope gradient (Brooks 1992, Chapter 3). A review of the literature indicates peat soil hydraulic conductivities range from greater than 200×10^{-3} cm s^{-1} in upper bog layers in Minnesota (Gafni and Brooks 1990) to 0.00011×10^{-3} cm s^{-1} in basal blanket peat in England (Holden and Burt 2003, Kolka et al. 2011). Bulk density varied from 0.8 g cm^{-3} in older lower layers of a Norwegian bog (Ohlson and Okland 1998) to 0.02 g cm^{-3} in the surface layer of a raised bog in New Brunswick (Korpijaako and Radforth 1972; Kolka et al. 2011).

These properties of peats have important ecological and economic consequences. The water table is often far below the surface in many peatlands, particularly in bogs, during the growing season (Boelter and Verry 1977; Bridgham and Richardson 1993; Verry 1997), and desiccation is an important constraint on the growth of *Sphagnum* mosses (Titus and Wagner 1984; Rydin 1985; Weltzin et al. 2001). Under drought conditions with a water table far below the surface, more decomposed peats would maintain higher plant-available water and faster transport of water to the roots (Päivänen 1973; Silins and Rothwell 1998).

Water retention and hydraulic conductivity are also important considerations in runoff from peatlands, drainage operations, and in commercial forestry in peatlands (Boelter 1964; Boelter and Verry 1977; Silins and Rothwell 1998). Drainage of highly decomposed, subsurface peats is quite difficult. Often effective drainage only occurs within 10 m or less of ditches (Bradof 1992a). As an example, failed attempts at draining the large Red Lake peatland complex in northwestern Minnesota from 1907 through the 1930s resulted in virtual bankruptcy of several counties and were only resolved when the state took over large areas of tax-delinquent lands (Bradof 1992b).

We have presented the traditional view of peatland hydrology. However, the work of Siegel and colleagues (Chason and Siegel 1986; Siegel 1988, 1992; Siegel and Glaser 1987; Glaser et al. 2004) has questioned the assumption that vertical flow is negligible in peatlands, and particularly in bogs because of very low conductivities in deep peat. With both field work and hydrologic modeling studies, they have demonstrated that the hydraulic head in raised bogs is sufficient to drive downward water flowpaths, making bogs recharge zones and adjacent fens discharge zones (see Chapter 3 for a discussion of these concepts). Even more interestingly, they have shown some bogs and fens to vary seasonally between being recharge and discharge zones. Chason and Siegel (1986) found much higher hydraulic conductivities in deep, decomposed peats than previous studies, which they attribute to discontinuous zones of buried wood, roots, and other structural features in peat that form "pipes" with extremely high conductivities. Working with the same group of scientists, Reeve et al. (2000) modeled vertical flow in peatlands. They found that vertical flow is negligible in raised bogs. Also, they determined that the amount of vertical flow depends on the differences in hydraulic conductivity with depth, especially at the catotelm/mineral soil boundary. Vertical flow can be more important when the mineral layer below the bog is permeable. If underlying sediment is impermeable, horizontal flow dominates. Further modelings work by Reeve et al. (2006) indicate that seasonal changes in water storage can influence the amount of vertical flow with high water tables with more head leading to higher vertical flows such as found during spring following snowmelt.

Runoff from peatlands outside permafrost areas is low, although it is higher in fens than bogs because of relatively constant groundwater inputs into fens and the potential for fens to also be present on gentle slopes (Boelter and Verry 1977; Verry 1997). However if permafrost is present, the infiltration and surface storage is low, and runoff occurs (Kane and Hinzman 1988). Free water mainly drains laterally above the permafrost following the slope. According to a study conducted in the interior of northeastern Russia, the ratio of water drained laterally to vertically is 8:1 (Alfimov and Ping 1994).

Peat Biogeochemistry: A Comparative Approach

Conterminous U.S. Peats: The Ombrogenous–Minerogenous Gradient

We examined 39 physical and chemical properties of soils from 20 different wetlands (Tables 10.3 and 10.4), 17 in northern Minnesota, 2 in North Carolina, and 1 in Florida. The Minnesota sites were part of a larger study in carbon and nutrient dynamics in wetlands and were placed along an ombrogenous–minerogenous gradient according to dominant vegetation and soil pH (Bridgham et al. 1998). While this gradient is strictly defined based on hydrology, field data generally show a close correspondence between hydrologic status, vegetation, and soil chemistry (Sjörs 1950; Heinselman 1963, 1970; Glaser 1987; Grootjans et al. 1988; Vitt and Chee 1990; Gorham and Janssens 1992; Vitt 2006). All sites were classified as Histosols, except for two of the Minnesota sites, Upper and Lower Shoepack, which were beaver meadows in Voyageurs National Park with a surface histic epipedon of from 8 to 21 cm thickness over a mineral layer.

The short pocosin (an ombrotrophic bog dominated by stunted ericaceous shrubs) and gum swamp (minerogenous forested swamp dominated by *Nyssa sylvatica*, *Liquidambar styraciflua*, *Acer rubrum*, and *Taxodium distichum*) sites in the Coastal Plain of North Carolina are described in Bridgham and Richardson (1993). The Florida Everglades site is dominated by sawgrass, *Cladium jamaicensis*. It is part of Water Conservation Area 2A and has not been impacted by agricultural runoff (C. Richardson, Duke University, personal communication). Five replicate cores from 0 to 25 cm depth were taken from hollows in each site, when significant microtopography was present.

We put the 39 variables from all 20 wetlands in Tables 10.3 and 10.4 into a principal component analysis (PCA; Wilkinson et al. 1992). PCA is a multivariate technique that combines the physical and chemical factors into master variables called components that explain the most variation in the data set. The correlation of all 39 variables with the three most important principal components is presented in Figure 10.4. The first principal component had high positive weightings from lignin, the lignin:cellulose ratio, bulk density, and the von Post index. In contrast, variables with high negative principal component 1 weightings were pyrophosphate color, rubbed and unrubbed fiber, water and acid soluble components, soluble phenolics, and extractable potassium. These variables suggest that principal component 1 describes a decomposition axis, with peat that has high positive values being highly decomposed.

The second principal component describes an alkalinity/pH axis, with high weightings from extractable Ca and Mg, the Ca:Mg ratio, cation-exchange capacity, total exchangeable bases, %base saturation, and pH (Figure 10.4). Interestingly, %humin, total soil nitrogen, and calcium-chloride extractable N clumped with these alkalinity variables, which suggest

TABLE 10.3

Chemical Characteristics of the Average of Five 0 to 25 cm Depth Cores from 20 Sites

Site	Type	Lat. N	Total Org. C%	Total N%	Total P%	C/N	C/P	N/P	%AFDM		
									Nonpolar Extr.	Water Soluble	Acid Soluble
Arlberg	Bog	46° 55′	41.7	1.26	0.044	33.2	971	29.0	7.00	11.60	45.9
Ash River	Bog	48° 24′	44.2	1.13	0.054	39.7	826	20.9	8.41	8.80	53.7
Pine Island	Bog	48° 17′	43.5	1.14	0.055	38.4	849	21.7	8.70	11.42	50.2
Red Lake	Bog	48° 22′	42.7	1.05	0.063	41.2	692	16.8	8.59	10.83	55.4
Toivola	Bog	47° 4′	42.2	1.12	0.039	37.8	1228	31.7	7.01	8.05	55.1
Marcell	Acidic fen	47° 31′	41.6	1.63	0.048	25.6	876	34.4	8.73	10.38	49.5
McGregor	Acidic fen	47° 39′	42.7	1.39	0.077	31.0	583	18.6	8.20	7.14	65.3
Alborn	Int. fen	47° 00′	38.9	2.61	0.082	14.9	479	32.1	4.75	7.31	48.3
Red Lake	Int. fen	48° 22′	43.3	2.43	0.076	17.9	573	32.0	7.66	7.13	28.8
Ash River	Tamar. sw.	48° 24′	42.9	1.98	0.078	21.8	558	25.6	8.56	6.83	28.3
Meadowlands	Tamar. sw.	47° 4′	42.6	2.57	0.114	16.6	380	22.9	5.32	5.52	39.1
Ash River	Cedar sw.	48° 24′	42.4	1.88	0.077	22.8	791	34.9	7.56	7.42	34.3
Isabella	Cedar sw.	47° 36′	42.6	1.85	0.076	23.3	584	25.0	6.20	8.92	30.5
Meadowlands	Cedar sw.	47° 3′	44.0	2.04	0.079	21.8	574	26.1	7.25	5.64	36.2
Gnesen	Ash sw.	46° 59′	34.7	2.58	0.266	13.5	131	9.7	6.16	8.27	37.0
Voyageurs	Meadow	48° 28′	21.1	1.53	0.127	13.7	166	12.0	6.83	7.22	35.6
Voyageurs	Meadow	48° 28′	23.8	1.59	0.115	15.0	212	13.9	6.57	9.05	30.8
North Carolina	Short pocosin	34° 55′	53.3	1.74	0.029	30.7	1838	59.8	7.25	4.97	25.1
North Carolina	Gum sw.	34° 55′	28.1	1.37	0.126	20.6	232	11.1	7.48	5.49	18.0
Florida	Everglades	26° 30′	45.0	2.99	0.034	15.1	1318	87.7	7.58	5.68	19.6

(*Continued*)

TABLE 10.3 (*Continued*)

Chemical Characteristics of the Average of Five 0 to 25 cm Depth Cores from 20 Sites

		%AFDM								
Site	**Type**	**Wat. Sol. Carbo.**	**Acid Sol. Carbo.**	**Soluble Phenolics**	**Lignin**	**Lignin/N**	**Lignin/ Cellulose**	**Mineral Content (%)**	**Bulk Density (mg m^{-3})**	**pH Water**
Arlberg	Bog	7.67	31.8	0.515	33.3	26.4	0.420	12.1	0.036	3.74
Ash River	Bog	7.89	39.4	0.568	29.8	26.7	0.356	5.2	0.050	3.75
Pine Island	Bog	7.38	35.7	0.649	28.4	24.8	0.361	6.3	0.054	3.80
Red Lake	Bog	7.30	41.7	0.687	24.1	22.9	0.304	6.0	0.052	3.72
Toivola	Bog	7.73	47.8	0.584	28.6	25.6	0.341	8.4	0.030	3.81
Marcell	Acidic fen	4.95	25.0	0.338	29.3	18.1	0.373	11.2	0.020	4.22
McGregor	Acidic fen	4.13	62.6	0.480	18.6	13.3	0.222	5.0	0.018	3.95
Alborn	Int. fen	3.84	14.5	0.203	38.2	14.6	0.441	22.7	0.100	4.84
Red Lake	Int. fen	6.64	24.3	0.215	53.3	22.0	0.650	12.5	0.076	5.57
Ash River	Tamar. sw.	5.66	15.5	0.291	54.3	27.6	0.658	13.4	0.102	5.91
Meadowlands	Tamar. sw.	2.58	17.6	0.247	48.7	19.1	0.555	13.4	0.104	5.63
Ash River	Cedar sw.	9.57	19.1	0.354	48.7	26.2	0.586	13.3	0.095	4.35
Isabella	Cedar sw.	2.15	7.7	0.155	52.2	28.5	0.632	14.7	0.147	6.61
Meadowlands	Cedar sw.	1.53	21.7	0.213	49.3	24.4	0.576	13.0	0.110	5.76
Gnesen	Ash sw.	3.42	8.7	0.096	44.5	17.3	0.546	29.3	0.150	6.13
Voyageurs	Meadow	3.05	5.5	0.070	43.0	28.9	0.548	55.4	0.213	5.75
Voyageurs	Meadow	5.06	4.1	0.070	45.9	28.9	0.598	52.3	0.236	6.16
North Carolina	Short pocosin	3.13	9.0	0.141	62.2	35.8	0.712	3.8	0.087	3.36
North Carolina	Gum sw.	2.43	3.6	0.019	63.2	47.0	0.778	46.1	0.206	3.97
Florida	Everglades	1.69	11.7	0.097	65.1	21.8	0.770	14.8	0.100	6.60

(*Continued*)

TABLE 10.3 (*Continued*)

Chemical Characteristics of the Average of Five 0 to 25 cm Depth Cores from 20 Sites

		cmolc/kg									
					Extractable Bases						
Site	Type	Exch. Acidity	Exch. Bases	CECpH 7	Na	K	Mg	Ca	Ca/Mg	Base Sat. (%)	Acid-F Extr. P. ($\mu g\ g^{-1}$)
Arlberg	Bog	19.9	13.2	33.1	0.382	2.87	3.3	6.7	2.01	39.6	3.26
Ash River	Bog	13.9	11.8	25.7	0.433	1.55	3.5	6.3	1.88	45.1	3.08
Pine Island	Bog	15.7	19.1	34.7	0.544	3.63	4.2	10.7	2.51	56.1	2.68
Red Lake	Bog	17.2	16.5	33.7	0.430	2.57	4.7	8.8	1.89	49.8	3.00
Toivola	Bog	24.0	18.1	42.2	0.465	2.83	3.9	10.9	2.82	43.2	1.12
Marcell	Acidic fen	31.3	21.1	52.4	0.550	1.65	5.0	13.9	2.79	40.8	1.94
McGregor	Acidic fen	30.7	22.0	52.6	0.793	4.42	5.7	11.1	1.97	41.8	36.91
Alborn	Int. fen	11.3	12.3	23.6	0.500	0.90	2.6	8.3	3.29	51.4	1.85
Red Lake	Int. fen	6.4	34.8	41.2	0.465	1.25	8.0	25.1	3.15	85.2	1.45
Ash River	Tamar. sw.	15.4	84.3	99.7	0.344	1.05	21.6	61.3	2.86	84.5	1.32
Meadowlands	Tamar. sw.	14.6	60.2	74.8	0.537	1.01	17.0	41.7	2.45	80.5	13.03
Ash River	Cedar sw.	19.6	31.9	51.5	0.422	1.55	6.8	23.1	3.45	62.8	4.91
Isabella	Cedar sw.	7.8	116.9	124.7	0.524	0.78	24.0	91.6	3.84	93.7	1.58
Meadowlands	Cedar sw.	19.1	67.0	86.1	0.381	0.70	17.4	48.5	2.78	77.8	1.80
Gnesen	Ash sw.	10.6	51.3	61.8	0.679	0.69	10.6	39.3	3.70	82.8	8.49
Voyageurs	Meadow	4.4	26.3	30.7	0.263	1.34	7.5	17.3	2.38	85.5	7.23
Voyageurs	Meadow	3.5	35.2	38.7	0.267	1.13	10.1	23.7	2.45	88.4	3.85
North Carolina	Short pocosin	28.1	7.1	35.2	0.733	0.68	5.3	0.4	0.07	20.1	0.82
North Carolina	Gum sw.	19.3	1.4	20.7	0.294	0.40	0.4	0.3	0.73	6.8	6.56
Florida	Everglades	5.5	139.9	145.4	7.658	1.94	30.3	100.0	3.30	96.2	8.01

(*Continued*)

TABLE 10.3 (*Continued*)

Chemical Characteristics of the Average of Five 0 to 25 cm Depth Cores from 20 Sites

		$CaCl_2$ Ext.		Oxalate Ext.		%AFDM		
Site	**Type**	**P (μg g^{-1})**	**N (μg g^{-1})**	**Fe (mg g^{-1})**	**Al (mg g^{-1})**	**Humin**	**Fulvic Acids**	**Humic Acids**
Arlberg	Bog	5.66	24.9	2.85	1.88	72.4	20.7	12.0
Ash River	Bog	0.96	0.7	0.90	0.46	72.1	0.6	23.1
Pine Island	Bog	6.16	14.6	0.93	0.57	72.7	23.8	7.6
Red Lake	Bog	11.95	9.6	2.61	1.01	72.3	3.5	21.3
Toivola	Bog	0.43	14.2	3.35	1.08	72.2	6.5	23.4
Marcell	Acidic fen	7.87	72.1	2.95	1.99	75.7	22.9	10.2
McGregor	Acidic fen	36.16	31.6	3.29	0.61	73.8	21.2	7.5
Alborn	Int. fen	0.53	16.1	12.29	3.24	75.9	26.5	23.0
Red Lake	Int. fen	0.91	26.6	6.48	2.08	79.8	14.7	13.8
Ash River	Tamar. sw.	2.38	85.1	2.76	1.63	79.2	3.3	11.5
Meadowlands	Tamar. sw.	1.79	39.0	7.34	1.19	70.6	4.8	11.7
Ash River	Cedar sw.	6.31	41.0	2.98	2.08	80.0	24.2	13.5
Isabella	Cedar sw.	1.02	48.6	6.09	1.09	80.2	0.9	13.4
Meadowlands	Cedar sw.	1.76	54.7	7.88	1.43	79.7	6.3	11.7
Gnesen	Ash sw.	0.20	42.2	10.48	5.61	77.1	13.5	17.2
Voyageurs	Meadow	3.09	19.8	5.70	3.03	73.2	57.0	26.4
Voyageurs	Meadow	0.80	38.9	6.29	2.39	63.5	54.0	23.2
North Carolina	Short pocosin	16.25	5.3	0.62	0.99	72.0	17.6	26.8
North Carolina	Gum sw.	0.41	7.0	1.38	7.64	54.4	42.6	37.1
Florida	Everglades	11.49	35.6	1.71	0.83	77.3	22.5	8.9

Note: All sites are in Minnesota except where noted. Int.: intermediate, Tamar.: tamarack, sw.: swamp, Lat.: latitude, Org.: organic, Extr.: extractable, AFDM: ash-free dry mass, Wat.: water, Sol.: soluble, Carbo.: carbohydrates, Exch.: exchangeable, CEC: cation exchange capacity, Sat.: saturation, Acid-F: acid fluoride.

TABLE 10.4

Physical Characteristics and Various Classification Schemes for the Average of Five 0 to 25 cm Depth Cores from 20 Sites

		Unrubbed Fiber		Rubbed Fiber		10YR				Classifications[a]			
Site	Type	% Dry-mass	% Volume	% Dry-mass	% Volume	Value Color	Chroma Color	Composite[b] Color	von Post Index	U.S.	Canadian	ASTM	von Post
Arlberg	Bog	78	73	68	53	7.2	2.2	5	3	2.4	2.6	2.8	2.8
Ash River	Bog	92	82	80	62	8	1.4	6.6	2.6	3	3	3	3
Pine Island	Bog	81	75	68	53	7.8	1.8	6	2.4	2.8	2.8	2.8	3
Red Lake	Bog	86	78	78	60	8	1.6	6.4	3.6	3	3	2.8	2.4
Toivola	Bog	78	73	74	57	7.8	1.8	6	2.8	2.8	2.8	2.8	3
Marcell	Acidic fen	76	72	60	48	8	1.4	6.6	4.6	3	3	3	2
McGregor	Acidic fen	85	78	77	60	8	1	7	2.8	3	3	3	3
Alborn	Int. fen	33	46	26	26	6.2	3.4	2.8	4.6	1.2	2	1.8	2
Red Lake	Int. fen	52	56	41	34	8	2	6	4	2	2	2	2
Ash River	Tamar. sw.	63	63	51	41	7.8	2.2	5.6	4.2	2.4	2.4	2.4	2
Meadowlands	Tamar. sw.	48	53	30	26	7	3.2	3.8	6.8	1.8	2	2	1.2
Ash River	Cedar sw.	53	57	44	36	7.6	2.4	5.2	6.2	2.4	2.4	2	1.6
Isabella	Cedar sw.	47	52	30	26	7	2.8	4.2	8	1.8	2	2	1
Meadowlands	Cedar sw.	58	60	43	35	6.8	2.8	4	6	2	2.2	2.2	1.6
Gnesen	Ash sw.	45	51	11	13	5.2	3	2.2	9.4	1	2	2	1
Voyageurs	Meadow	31	42	20	19	6.8	3.2	3.6	6	1.6	2	1.4	2
Voyageurs	Meadow	35	45	17	17	5.6	3	2.6	6.4	1	2	1.6	1.4
North Carolina	Short pocosin	48	53	16	16	5	3	2	10	1	2	2	1
North Carolina	Gum sw.	28	40	8	10	3	2	1	10	1	1.4	1.2	1
Florida	Everglades	63	63	20	19	5.8	3	2.8	9	1	2	2.2	1

Note: All sites are in Minnesota except where noted. See text for description of classification schemes. Int.: intermediate, Tamar.: tamarack, sw.: swamp.

[a] Average classification for the five cores. 1 = sapric, 2 = hemic, 3 = fibric.

[b] 10YR Value—10YR Color (Parent and Caron 1993).

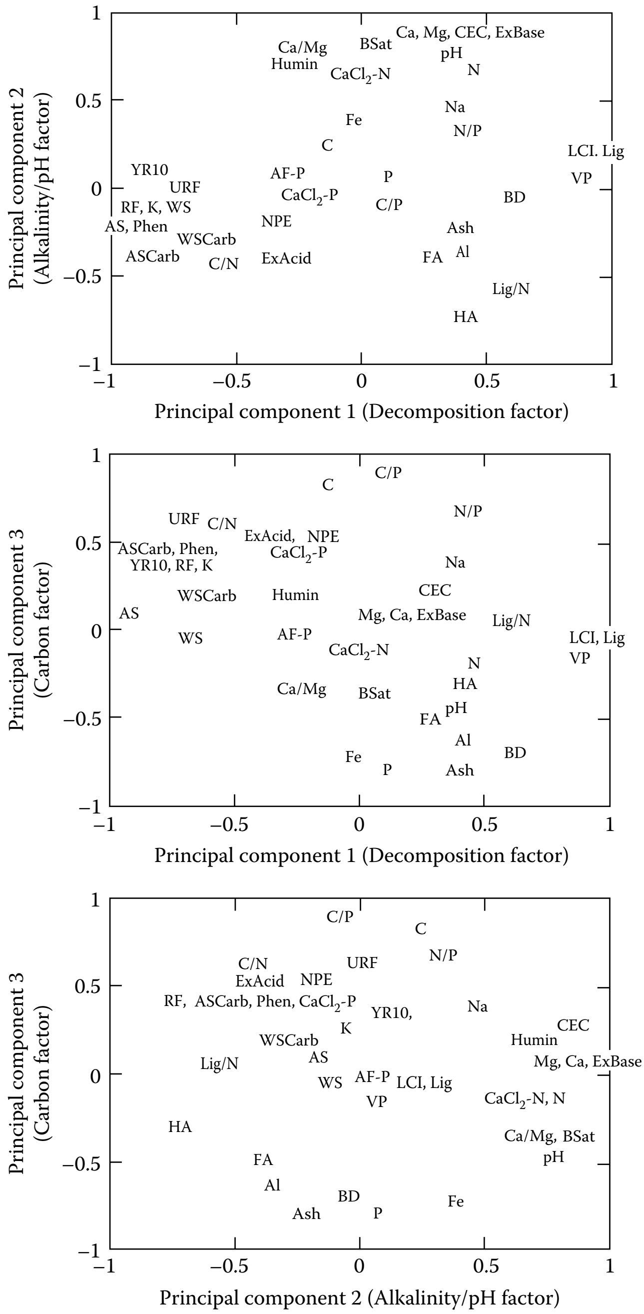


FIGURE 10.4
The loadings of 39 soil variables from 15 peatlands in northern Minnesota, 2 beaver meadows with histic epipedons in northern Minnesota, 2 peatlands in North Carolina, and 1 peatland in Florida on the first 3 axes of a principal components analysis. The loading is comparable to the correlation coefficient (*r*) for each variable against each axis. Abbreviations are as in Table 10.5.

a positive relationship between alkalinity/pH, humin formation, and nitrogen pools and fluxes. In contrast, humic acid content had a high negative weighting on this axis, which suggests it has a negative relationship with alkalinity/pH.

The third principal component axis was related to soil carbon and mineral content (Figure 10.4). It had positive weighting from total soil carbon, and high negative weightings from %mineral content, oxalate-extractable Fe and Al, bulk density, and total soil phosphorus. Phosphorus is strongly sorbed by iron and aluminum hydroxyoxides, so it is not surprising that greater mineral content is related to higher total soil phosphorus levels, although this does not necessarily translate into higher available phosphorus (Bridgham et al. 1998). Additionally, more minerogenous peats may receive greater inputs of apatite–phosphorus from weathering.

There is a large cost in labor, time, and expense in doing many of these chemical analyses, and it is promising that a simple set of physical and chemical variables often measured in peats is closely correlated with many of the more difficult chemical analyses. In particular, mineral content, bulk density, pH, fiber content, and the von Post index are correlated with many other chemical variables (Table 10.5). They are also as effective as the chemical variables in predicting nutrient and carbon mineralization in peats (Lévesque and Mathur 1979; Bridgham et al. 1998).

PCA also allows one to determine "factor scores" for each of the 20 wetlands along these three principal component axes. We used our multivariate data set to discriminate natural groupings of peatlands according to their soil characteristics (Figure 10.5). The first, second, and third factors explained 28.8%, 22.5% and 21.1%, respectively, of the variance among sites, or 72.4% of the total variance. The first factor (Decomposition Factor) effectively separated three groups of wetlands: acidic fens and bogs, more minerotrophic northern wetlands, and southern peatlands. The second factor (Alkalinity/pH Factor) separated bogs from acidic fens, beaver meadows from minerotrophic northern peatlands, and the alkaline Everglades site from the acidic North Carolina peatlands. The third factor (Carbon Factor) separated minerotrophic northern cedar and tamarack swamps from intermediate fens, the ash swamp site, and the two beaver meadows. The nutrient-deficient short pocosin and Everglades sites were separated from the relatively nutrient-rich North Carolina gum swamp.

The difficulty of applying the ombrogenous–minerogenous gradient to southern peatlands is evident from our data. One sees the expected decrease in rubbed fiber content and increase in mineral ash, lignin, pH, %base saturation, and related variables expressing increasing alkalinity from bogs to ash swamps and beaver meadows in the northern sites, related to increasing minerogenous water inputs and their impact on water chemistry (Table 10.3, Figure 10.6). However, both short pocosins and the Everglades are profoundly phosphorus limited, whereas the gum swamp is relatively fertile (Walbridge 1991; Koch and Reddy 1992; Bridgham and Richardson 1993; Craft and Richardson 1997). Hydrologically, the Everglades site would be considered a "poor" fen, despite its alkaline soil conditions, and the gum swamp is a highly minerogenous "rich" swamp forest, despite its very acidic soil (Table 10.3, Figure 10.6). The sands of the North Carolina Coastal Plain have very low exchangeable basic cation concentrations, so contribute little alkalinity despite being highly minerogenous (Bridgham and Richardson 1993). Additionally, all of the southern peats are highly decomposed hemic or sapric peats with very low fiber and cellulose content, but high lignin content (Tables 10.3 and 10.4; Figures 10.4 through 10.6).

Our data support the traditional concept of an ombrogenous–minerogenous gradient in northern peatlands in terms of alkalinity and degree of decomposition of peats; however, soil nutrient availability is more problematic. We found in these same Minnesota

TABLE 10.5

Pearson Correlations (r) when $P < 0.05$ for Variables in Tables 10.3 and 10.4

	C	N	P	C/N	C/P	N/P	NPE	WS	AS	WSCarb
C										
N										
P	−0.57									
C/N	0.48	−0.78	−0.55							
C/P	0.74		−0.74	0.51						
N/P	0.56	0.44	−0.55		0.77					
NPE		−0.52		0.58						
WS		−0.54		0.51						
AS		−0.49		0.63				0.55		
WSCarb		−0.51		0.59				0.59	0.46	
ASCarb		−0.51		0.75			0.47		0.85	0.52
Phen	0.44	−0.62	−0.46	0.88			0.50	0.63	0.80	0.73
Lig		0.58		−0.62				−0.69	−0.97	−0.52
Lig/N									−0.58	
LCI		0.54		−0.64				−0.63	−0.99	−0.49
Ash	−0.95		0.57	−0.64	−0.65					
BD	−0.77		0.58	−0.69	−0.58		−0.44		−0.66	−0.48
pH		0.69		−0.78	−0.45				−0.53	−0.54
ExAcid	0.47			0.55	0.46				0.44	
ExBase		0.60							−0.46	−0.47
CEC		0.55				0.48				−0.46
Na		0.50				0.80				
K		−0.49		0.62			0.45	0.50	0.71	0.47
Mg		0.60		−0.44					−0.48	−0.50
Ca		0.60		−0.44					−0.47	−0.47
Ca/Mg		0.51								
BSat		0.57		−0.56						
AF-P										
$CaCl_2$-P										
$CaCl_2$-N										
Fe		0.58	0.59	−0.63	−0.58		−0.77			
Al	−0.62		0.69	−0.49	−0.55					
Humin	0.51									
FA	−0.78									
HA	−0.47									
URF	0.56	−0.48	−0.51	0.82	0.45		0.68	0.52	0.69	0.55
RF		−0.60	−0.51	0.83			0.59	0.57	0.81	0.67
YR10		−0.45	−0.45	0.63			0.62	0.52	0.70	0.65
VP		0.49		−0.59				−0.58	−0.78	−0.68

(*Continued*)

TABLE 10.5 (*Continued*)

Pearson Correlations (*r*) when $P < 0.05$ for Variables in Tables 10.3 and 10.4

	ASCarb	Phen	Lig	Lig/N	LCI	Ash	BD	pH	ExAcid	ExBase
C										
N										
P										
C/N										
C/P										
N/P										
NPE										
WS										
AS										
WSCarb										
ASCarb										
Phen	0.87									
Lig	−0.81	−0.80								
Lig/N			0.52							
LCI	−0.84	−0.81	0.99	0.55						
Ash	−0.64	−0.66								
BD	−0.81	−0.78	0.58	0.49	0.64	0.90				
pH	−0.59	−0.65	0.51		0.53		0.57			
ExAcid	0.53					−0.53	−0.63	−0.73		
ExBase			0.49		0.47			0.81	−0.45	
CEC			0.44					0.70		0.98
Na										0.63
K	0.84	0.76	−0.74		−0.73		−0.64	−0.48		
Mg			0.52		0.50			0.81		0.98
Ca			0.50		0.48			0.82	−0.47	1.00
Ca/Mg				−0.58				0.64	−0.45	0.57
BSat								0.90	−0.74	0.76
AF-P										
$CaCl_2$-P	0.53						−0.44		0.54	
$CaCl_2$-N								0.56		0.56
Fe				−0.48				0.55		
Al	−0.53	−0.56				0.68	0.60			
Humin				−0.60		−0.51				0.48
FA		−0.46				0.81	0.61			
HA				0.66		0.54	0.51			−0.49
URF	0.86	0.88	−0.68		−0.70	−0.75	−0.85	−0.51	0.48	
RF	0.91	0.95	−0.80		−0.81	−0.66	−0.81	−0.57	0.47	
YR10	0.81	0.81	−0.71		−0.71	−0.61	−0.75			
VP	−0.79	−0.82	0.80		0.80		0.63			

(*Continued*)

TABLE 10.5 (*Continued*)

Pearson Correlations (*r*) when $P < 0.05$ for Variables in Tables 10.3 and 10.4

	CEC	Na	K	Mg	Ca	Ca/Mg	BSat	AF-P	$CaCl_2$-P	$CaCl_2$-N
C										
N										
P										
C/N										
C/P										
N/P										
NPE										
WS										
AS										
WSCarb										
ASCarb										
Phen										
Lig										
Lig/N										
LCI										
Ash										
BD										
pH										
ExAcid										
ExBase										
CEC										
Na	0.62									
K										
Mg	0.96	0.59								
Ca	0.97	0.59		0.97						
Ca/Mg	0.51			0.48	0.60					
BSat	0.64			0.76	0.76	0.74				
AF-P			0.49							
$CaCl_2$-P			0.63					0.75		
$CaCl_2$-N	0.61			0.59	0.57	0.53	0.51			
Fe						0.55	0.46			
Al			−0.49							
Humin	0.50			0.45	0.48	0.67	0.53			0.47
FA										
HA	−0.55			−0.47	−0.47	−0.53	−0.45			−0.58
URF			0.74							−0.49
RF			0.75							−0.49
YR10			0.66							
VP			−0.68							

(*Continued*)

TABLE 10.5 (*Continued*)

Pearson Correlations (r) when $P < 0.05$ for Variables in Tables 10.3 and 10.4

	Fe	Al	Humin	FA	HA	URF	RF	YR10
C								
N								
P								
C/N								
C/P								
N/P								
NPE								
WS								
AS								
WSCarb								
ASCarb								
Phen								
Lig								
Lig/N								
LCI								
Ash								
BD								
pH								
ExAcid								
ExBase								
CEC								
Na								
K								
Mg								
Ca								
Ca/Mg								
BSat								
AF-P								
$CaCl_2$-P								
$CaCl_2$-N								
Fe								
Al								
Humin		−0.49						
FA		0.47	−0.51					
HA		0.57	−0.67					
URF	−0.53	−0.64		−0.54	−0.47			
RF		−0.60		−0.47		0.92		
YR10		−0.63		−0.44	−0.52	0.83	0.92	
VP		0.52				−0.70	−0.88	−0.85

Note: C: %organic C, N: %total N, P: %total P, NPE: nonpolar extractable organic matter, WS: water soluble organic matter, AS: acid soluble organic matter, WSCarb: water soluble carbohydrates, ASCarb: acid soluble carbohydrates, Phen: soluble phenolics, Lig: lignin, LCI: lignin/cellulose, BD: bulk density, ExAcid: exchangeable acidity, ExBase: exchangeable bases, CEC: cation-exchange capacity, BSat: %base saturation, AF-P: acid fluoride extractable P, $CaCl_2$-N and $CaCl_2$-P: calcium chloride extractable N and P, FA: fulvic acid, HA: humic acid, URF: %unrubbed fiber, RF: %rubbed fiber, YR10: composite pyrophosphate color (10YR Value—10YR Color), VP: von Post index.

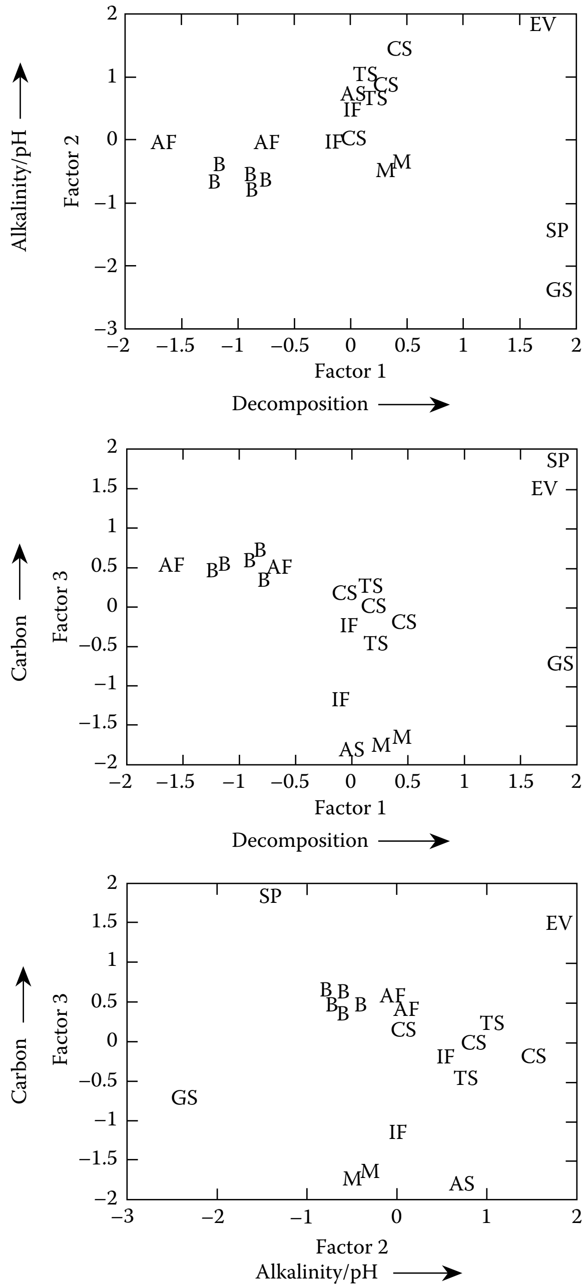


FIGURE 10.5
Factor scores for the first three axes of a principal components analysis of 39 soil variables from 15 peatlands in northern Minnesota, 2 beaver meadows with histic epipedons in northern Minnesota, 2 peatlands in North Carolina, and 1 peatland in Florida. B: bog, AF: acidic (poor) fen, IF: intermediate fen, TS: tamarack swamp, CS: cedar swamp, AS: black ash swamp, M: beaver meadow, SP: short pocosin (NC), GS: gum swamp (NC), and EV: Everglades (FL). The northern wetlands occur across an ombrogenous–minerogenous gradient in the order listed above, whereas hydrologically SP is an ombrogenous bog, GS is a minerogenous swamp forest, and EV is a "poor" fen.

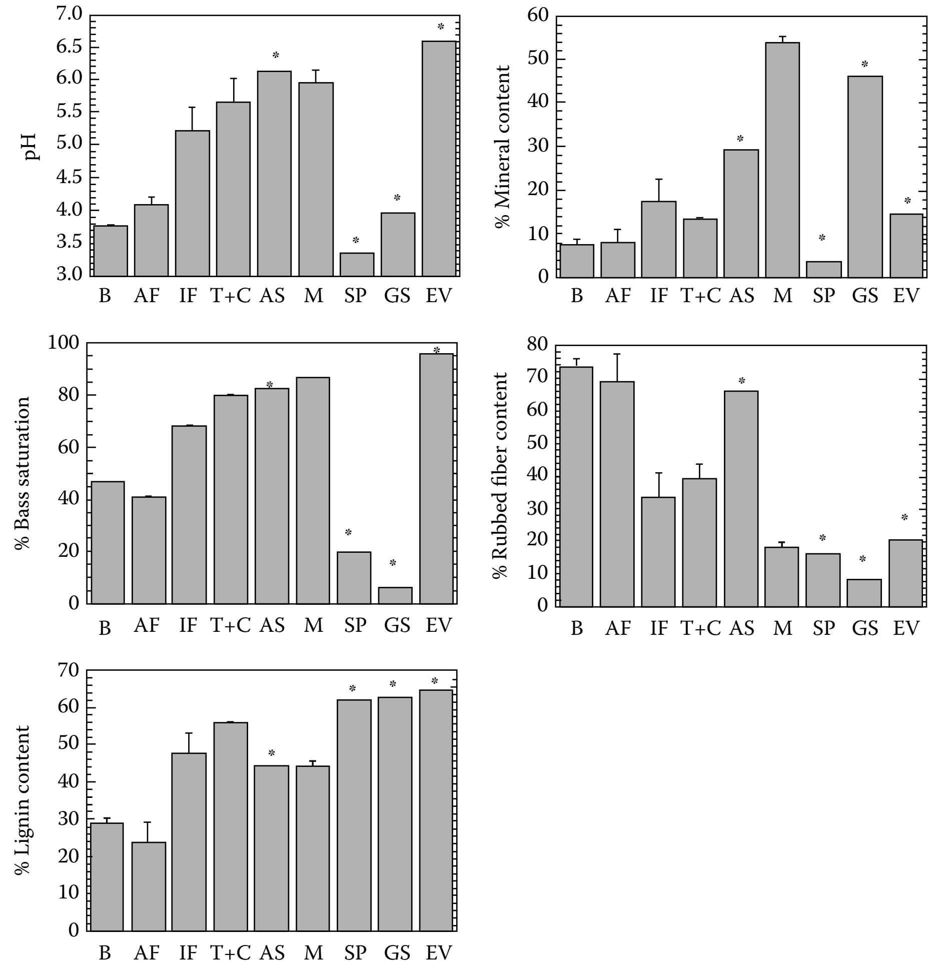


FIGURE 10.6
Relationship between pH, mineral content, base saturation, rubbed fiber content, and lignin to the ombrogenous–minerogenous gradient in the northern wetlands (going from left to right on the *x*-axis) and the three southern peatlands (SP, GS, and EV). The northern wetlands occur across an ombrogenous–minerogenous gradient in the order listed above, whereas hydrologically SP is an ombrogenous bog, GS is a minerogenous swamp forest, and EV is a "poor" fen. Average ± 1 standard error, except * indicates $N = 1$ site so standard errors could not be obtained.

wetlands that more minerogenous wetlands have larger total soil nitrogen and phosphorus pools, but those pools turn over more slowly in minerogenous sites (Bridgham et al. 1998). A phosphorus isotope addition experiment across the ombrogenous–minerogenous gradient resulted in no differences in available phosphorus, microbial phosphorus, and the root phosphorus at 10–20 cm, although total soil phosphorus and aboveground vegetation phosphorus content increased from bog to rich fen (Kellogg and Bridgham 2003). It appears that although bogs and intermediate fens have a small total phosphorus pool, they have similar phosphorus availability to rich fens because of rapid cycling and efficient retention of phosphorus. The large increase in bulk density in more minerogenous sites also has important consequences, because plant roots and microbes exploit a volume and not a mass of soil. The net result of all these factors was that nitrogen availability was higher in more minerogenous Minnesota wetlands (Bridgham et al. 1998). Chapin (1998) conducted a detailed fertilization experiment in an intermediate fen and bog in northern Minnesota and found similar results. Interestingly, she found that bog vegetation was not nutrient limited, except for a delayed response in ericaceous shrubs, and *Sphagnum* mosses were actually inhibited at moderate rates of nitrogen addition. The fen vegetation was phosphorus limited. Similar results have been found for both soil nutrient availability (Waughman 1980; Verhoeven et al. 1990; Koerselman et al. 1993; Updegraff et al. 1995) and plant-nutrient response (Clymo 1987; Lee et al. 1987; Boyer and Wheeler 1989; Bridgham et al. 1996) in other northern peatlands. More recent research using enzymes to determine nutrient limitations found phosphorus to be more limiting than nitrogen across a gradient of ombrogenous to minerogenous peatlands in northern Minnesota (Hill et al. 2014).

We suggest that the ombrogenous–minerogenous paradigm is an important and useful concept in northern peatlands, although its relation to a nutrient availability gradient appears to be complicated and worthy of further research. We conclude that the ombrogenous–minerogenous gradient does not appear to be directly translatable into an oligotrophic–eutrophic gradient. Furthermore, traditional concepts of how the ombrogenous-minerogenous gradient affects peat chemistry and physical properties in northern peatlands do not appear to be useful in southern peatlands.

Alaskan Peatlands: Histosols and Gelisols

We also examined a more limited set of soil variables in peats collected in the five pedons from Alaska (Table 10.6). Pedons 1 and 2 are intermediate fens, whereas pedons 3 through 5 are bogs. Pedons 3 and 5 are Histosols, whereas pedons 1, 2, and 4 are Histels within the order Gelisols.

The bulk density and mineral content of the horizons from the five pedons from Alaska are much higher than those from the Minnesota and southern peats (Table 10.3). Eolian and volcanic deposits (loess and tephra) have been active in many parts of Alaska and northwest Canada since the Late Pleistocene (Péwé 1975; Riehle 1985). Because of this frequent or intermittent input of mineral deposits, the organic soils in these regions have a higher bulk density compared with those developed in the humid maritime zones of southeastern Alaska and British Columbia. The additions of these materials appear in bands and layers in the peat, and thus they can serve as time-stratigraphic markers. In peat developed in bottom lands, mineral layers exist in lamella or bands due to the erosion or washing from surrounding slopes (Pedon 2).

In northern Alaska, as in the Minnesota sites, vegetation and land cover class show a strong correlation with the base status and pH of the soil (Ping et al. 1998). The pH of Alaskan peatlands decreases from 5.5 to 7.7 in the Arctic coast to 4.0 to 4.5 in the boreal forest in the interior, to 3.0 to 3.5 in south central and southeastern Alaska. Most bogs in south central Alaska are extremely acidic and have low base status. Some of the bogs have hydraulic conductivity less than 10 cm h^{-1} (Clark and Kautz 1997). Péwé (1975) pointed out that there is continuous deposition of carbonate-rich loess in the Arctic Coastal Plain and in interior Alaska if streams are transporting glacial debris. In these soils, extractable Ca and Mg dominate the soluble salts and the exchange sites in the soils (Pedons 1 and 2). Pedon 3 is a raised bog with *Sphagnum* moss as the dominant vegetation making the pH very acidic. Even though the area has relatively low loess deposition, the added carbonates from the loess are reflected in the Ca-dominance of the exchange sites and the slightly higher base saturation in the surface layer. Although Pedon 4 formed in humid

TABLE 10.6

Characteristics of Selected Histosols from Alaska

Pedon #Lat. N	Horizon	Depth (cm)	Total Org. C (%)	C/N	Mineral Content (%)	Bulk Density (mg m^{-3})	pH $CaCl_2$	cmolc/kg Exch. Acidity	CEC	Extractable Bases Na	K	Mg	Ca	Base Sat. (%)	Fiber Content Unrub. (%)	Rubbed (%)	Pyrophosphate Color
1	Oa1	0–18	23	35	62	0.39	6.9	17	79	1	tr	4	85	100	52	16	10YR 5/3
70° 17′	Oa2	18–39	15	13	77	0.49	5.9	13	29	1	tr	1	16	61	26	12	10YR 4/3
	Oe	39–50	22	14	65	0.38	6.3	18	49	1	tr	4	43	99	58	24	10YR 4/3
	Oef	50–100	25	19	61	n.d.	7.1	12	63	2	tr	6	115	100	80	26	10YR 4/3
	Cf	39–80	tr	11	n.d.	1.8	7.7	n.d.	2	0	tr	1	n.d.	100	n.d.	n.d.	n.d.
2	Oi	0–17	51	25	15		7.7	21	168	tr	3	20	183	100	92	64	10YR 7/3
67° 26′	Oe1	17–35	49	17	20	0.12	6.7	38	197	tr	tr	15	186	100	64	36	10YR 6/3
	Oe2	35–48	52	29	n.d.		5.8	49	160	tr	tr	11	146	98	n.d.	n.d.	n.d.
	C/Oa	48–54	24	n.d.	n.d.		5.1	47	89	tr	tr	6	81	99	n.d.	n.d.	n.d.
	Oef1	54–85	42	21	32		5.4	66	139	tr	tr	6	113	86	n.d.	n.d.	n.d.
	Oaf	85–95	29	20	54		n.d.	58	109	tr	tr	5	90	87	52	16	10YR 3/3
	Oef2	95–108	n.d.	n.d.	87		5.6	18	25	tr	tr	1	22	94	20	12	10YR 5/3
3	Oe	0–31	39	19	15	0.13	4.3		136	1	1	15	40	42	70	40	7.5YR7/5
64° 52′	Oi	31–61	38	31	9	0.1	3.9		116	tr	tr	5	12	16	88	75	10YR 8/2
	Oif	61–127	38	26	7	n.d.	4.4		81	tr	tr	5	12	22	90	80	10YR 8/1
4	Oi	0–29	53	68		n.d.	4.1	79	129	1	1	13	63	60			
61° 25′	Oe	29–47	53	32		0.15	4.3	83	106	1	tr	6	49	52			
	Oa	47–79	49	27		0.4	4.3	82	96	1	tr	4	42	49			
	O′e1	79–97	22	30		0.57	4.5	61	49	1	tr	2	18	40			
	O′e2	97–148	55	28		0.2	4.5	78	106	1	tr	4	48	50			
	O′I	148–165	60	32		n.d.	4.5	85	132	1	tr	6	68	57			
5	Oi	0–3			17		3.3	94	99	1	2	7	13	23	76	56	10YR 8/3
56° 30′	Oe	3–18			7		3	130	132	1	1	10	13	19	62	42	7.5YR 8/2
	Oa	18–94			54		3.5	72	76	1	1	2	4	9	48	30	5YR 3/4

south central Alaska, the base saturation is higher than that of Pedon 3 because it is on a broad flood plain which collects seasonal input of minerals. Pedon 5 is a well-drained Folist in perudic southeastern Alaska. Its soil is strongly acidic (pH at 3.3) and has very low base saturation.

Ping et al. (1997b) found that organic matter in fens of the arctic coast was dominated by cellulose (approximately 50%), whereas the humin fraction was <20%. Humic acids dominated the soluble fractions, and the C/N ratio ranged from 6 to 17. In comparison, the Minnesota peats had generally <40% cellulose (i.e., acid-soluble carbohydrates), >70% humin, a variable humic acid:fulvic acid ratio, and a C/N ratio which ranged from 14 to 41 (Table 10.3). All these data point to a lesser degree of humification of peats as the climate gets colder. This generalization is borne out by a similar comparison of the Minnesota peats to those in North Carolina and Florida in Table 10.3.

Peat formed in the zone of continuous permafrost, such as arctic Alaska and northwest Canada, contains cryogenic features such as ice lenses, ice wedges, and other types of ground ice, generally at a depth of 40–60 cm (Tarnocai et al. 1993; Ping et al. 1997a, b, 1998). The upper permafrost layer of these soils often contains up to 80% ice by volume. Cryoturbation causes mixing of soil horizons and redistribution of carbon, resulting in significant carbon stores in the permafrost (Michaelson et al. 1996; Tarnocai et al. 2009). Thawing as a result of climate change is predicted to have important positive feedbacks to the global carbon cycle thereby increasing warming potential (Schurr et al. 2008; Kovan et al. 2011).

Our emphasis in this comparative biogeochemical approach has been on the peatlands of the U.S. A multivariate analysis of numerous soil properties of Canadian bogs was performed by Brown et al. (1990), but their emphasis was not on the ombrogenous–minerogenous gradient, and the study was done within a more limited geographical setting. Additionally, a wealth of information on Canadian peats is found in National Wetlands Working Group (1988). The review by Clymo (1983) emphasizes European peatlands and has long been a classic in this field. Bohlin et al. (1989) examined a wide range of peat properties in a diverse group of Swedish peats and used principal components analysis to examine their results. They found that the peats were differentiated by botanical composition and degree of decomposition, and particularly emphasized the differences between *Sphagnum* (bog) and *Carex* (fen)-derived peats. *Carex* peats were more humified due to microbial decomposition than *Sphagnum* peats. A thorough review of humic substances in peats is provided by Mathur and Farnham (1985). Vitt (2006) used a five-factor approach integrating hydrology, climate, chemistry, substrate, and vegetation into a practical model to classify peatlands and natural gradients among peatland types. He developed functional levels of organization based on the five factors and used this framework to construct chronological "grades" that begin at wetland initiation followed by peatland development and then ultimately differentiate peatlands between bogs and fens (Vitt 2006).

Conclusions

We have presented here a framework for understanding the physical and biogeochemical properties of peat based upon the ombrogenous–minerogenous gradient, and examined how the properties of this gradient differ among different climatic zones. It is clear from the work presented here and elsewhere that peatlands are a critical ecosystem for many reasons. Peatlands harbor key flora and fauna, contribute to clean water, mitigate flooding, and store vast amounts of carbon. As a result of centuries of carbon accumulation, peatlands have mitigated rising concentrations of carbon dioxide in the atmosphere. Warming of Histosols and Gelisols will lead to positive feedbacks to the atmosphere, possibly accelerating climate change. Management approaches and research aimed at mitigating or adapting to climate change should be a priority for these globally important ecosystems.

Acknowledgments

We would like to thank Karen Updegraff and J. L. Richardson for their contributions to the first edition of this chapter, Curtis Richardson for obtaining the Everglades peat cores, Anastasia Bamford for technical assistance in the laboratory, John Pastor for comments on an earlier version of this manuscript, Inez Kettles for information on the extent of Canadian peatlands, and Sharon Waltman for access to the STATSGO and MUIR NRCS databases. This research was funded by a grant from NASA's Terrestrial Biosphere Program, the National Science Foundation (DEB-9496305, DEB-9707426), and a Distinguished Global Change Postdoctoral Fellowship from the Department of Energy to Scott Bridgham. The research in Alaska was supported by the USDA-Hatch program.

References

Ahrens, R. J., J. G. Bockheim, and C.L. Ping. 2004. Chapter 3: The Gelisol order in soil taxonomy. In J. M. Kimble (Ed.) *Cryosols* Springer, Berlin Heidelberg, pp. 627–635.

Alfimov, A. V. and C. L. Ping. 1994. Water regimes of boggy soils on slopes in the Upper Kolyma Basin, NE Russia. *Agronomy Abstr. Agron. Soc. Am.* Madison, WI.

Aselmann, I. and P. J. Crutzen. 1989. Global distribution of natural freshwater wetlands and rice paddies, their net primary productivity, seasonality and possible methane emissions. *J. Atmos. Chem.* 8: 307–359.

ASTM. 2013. *Annual Book of ASTM Standards*. Vol. 04.08. ASTM, Philadelphia, PA.

Baden, W. and R. Eggelsmann. 1963. Zur Durchlässigkeit der Moorboden. *Z. Kulturtech. Flurber.* 4: 226–254.

Belyea, L. R. and A. J. Baird. 2006. Beyond "the limits to peat bog growth": Cross-scale feedback in peatland development. *Ecol. Monogr.* 76: 299–322.

Boelter, D. H. 1964. Water storage characteristics of several peats *in situ. Soil Sci. Soc. Am. Proc.* 28: 433–435.

Boelter, D. H. 1968. Important physical properties of peat materials. In *Proc. 3rd Int. Peat Congress*. International Peat Society, Secretariat, Kuokkalantie, 4, FIN-40520, Jyväskylä, Finland, pp. 150–156.

Boelter, D. H. 1969. Physical properties of peats as related to degree of decomposition. *Soil Sci. Soc. Am. Proc.* 33: 606–609.

Boelter, D. H. and E. S. Verry. 1977. *Peatland and Water in the Northern Lake States.* USDA Forest Service General Technical Report NC31. North Central Forest Experimental Station, Forest Service, USDA, St. Paul, MN.

Bohlin, E., M. Hämäläinen, and T. Sundén. 1989. Botanical and chemical characterization of peat using multivariate methods. *Soil Sci.* 147: 252–263.

Botch, M. S., K. I. Kobak, T. S. Vinson, and T. P. Kolchugina. 1995. Carbon pools and accumulation in peatlands of the former Soviet Union. *Glob. Biogeochem. Cycles* 9: 37–46.
Boyer, M. L. H. and B. D. Wheeler. 1989. Vegetation patterns in spring-fed calcareous fens: Calcite precipitation and constraints on fertility. *J. Ecol.* 77: 597–609.
Bradof, K. L. 1992a. Ditching of the Red Lake Peatland during the homestead era. In H. E. Wright Jr., B. A. Coffin, and N. E. Aaseng (Eds.) *The Patterned Peatlands of Minnesota*. University of Minnesota Press, Minneapolis, MN, pp. 263–284.
Bradof, K. L. 1992b. Impact of ditching and road construction on Red Lake Peatland. In H. E. Wright Jr., B. A. Coffin, and N. E. Aaseng (Eds.) *The Patterned Peatlands of Minnesota.* University of Minnesota Press, Minneapolis, MN, pp. 173–186.
Brady, N.C. and R.R. Weil. 2008. *The Nature and Properties of Soils*, 14th ed. Pearson Prentice Hall, Upper Saddle River, NJ.
Bridgham, S. D., J. Pastor, J. Janssens, C. Chapin, and T. Malterer. 1996. Multiple limiting gradients in peatlands: A call for a new paradigm. *Wetlands* 16: 45–65.
Bridgham, S. D. and C. J. Richardson. 1993. Hydrology and nutrient gradients in North Carolina peatlands. *Wetlands* 13: 207–218.
Bridgham, S., J. Megonigal, J. Keller, N. Bliss, and C. Trettin. 2006. The carbon balance of North American wetlands. *Wetlands* 26(4): 889–916.
Bridgham, S. D., K. Updegraff, and J. Pastor. 1998. Carbon, nitrogen, and phosphorus mineralization in northern wetlands. *Ecology* 79: 1545–1561.
Brooks, K. N. 1992. Surface hydrology. In H. E. Wright Jr., B. A. Coffin, and N. E. Aaseng (Eds.) *The Patterned Peatlands of Minnesota.* University of Minnesota Press, Minneapolis, MN, pp. 153–162.
Brown, D. A., S. P. Mathur, A. Brown, and K. J. Kushner. 1990. Relationships between some properties of organic soils from the southern Canadian Shield. *Can. J. Soil Sci.* 90: 363–377.
Camill, P. and J. S. Clark. 1998. Climate change disequilibrium of boreal permafrost peatlands caused by local processes. *Am. Naturalist* 151: 207–222.
Chapin, C. T. 1998. *Plant community response and nutrient dynamics as a result of manipulations of pH and nutrients in a bog and fen in northeastern Minnesota*. PhD dissertation, University of Notre Dame, Notre Dame, IN.
Chason, D. B. and D. I. Siegel. 1986. Hydraulic conductivity and related physical properties of peat, Lost River peatland, northern Minnesota. *Soil Sci.* 142: 91–99.
Clark, M. H. and D. R. Kautz. 1997. *Soil Survey of Matanuska–Susita Valley Area, Alaska*. USDA Natural Resources Conservation Service. U.S. Govt. Printing Office. Washington, DC.
Clymo, R. S. 1983. Peat. In A. J. Gore (Ed.) *Mires: Swamp, Bog, Fen and Moor. Ecosystems of the World, 4A*. Elsevier Scientific Publishing, New York, NY, pp. 159–224.
Clymo, R. S. 1987. Interactions of *Sphagnum* with water and air. In T. C. Hutchinson and K. M. Meema (Eds.) *Effects of Atmospheric Pollutants on Forests, Wetlands and Agricultural Ecosystems.* Springer-Verlag, Berlin, Germany, pp. 513–529.
Cowardin, L. M., V. Carter, F. C. Golet, and E. T. LaRoe. 1979. *Classification of Wetlands and Deepwater Habitats of the United States*. FWS/OBS79/31, Fish and Wildlife Service, U.S. Department of the Interior, Washington, DC.
Craft, C. B. and C. J. Richardson. 1997. Relationships between soil nutrients and plant species composition in Everglades peatlands. *J. Environ. Qual.* 26: 224–232.
Damman, A. W. H. 1986. Hydrology, development, and biogeochemistry of ombrogenous peat bogs with special reference to nutrient relocation in a western Newfoundland bog. *Can. J. Bot.* 64: 384–394.
Damman, A. W. H. and T. W. French. 1987. *The Ecology of Peat Bogs of the Glaciated Northeastern United States: A Community Profile*. Fish and Wildlife Service Report 85 (7.16), U.S. Department of Interior, Washington, DC.
Day, J. H., P. J. Rennie, W. Stanek, and G. P. Raymond. 1979. *Peat Testing Manual.* Technical Memorandum Number 125, Associate Committee on Geotechnical Research. National Research Council of Canada, Ottawa, Canada.

Gafni, A. and K. N. Brooks. 1990. Hydraulic characteristics of four peatlands in Minnesota. *Can. J. Soil Sci.* 70: 239–253.
Glaser, P. H. 1987. *The Ecology of Patterned Boreal Peatlands of Northern Minnesota: A Community Profile.* Fish and Wildlife Service Report 85 (7.14), U.S. Department of Interior, Washington, DC.
Glaser, P. H., B. C. S. Hansen, D. I. Siegel, A. S. Reeve, and P. J. Morin. 2004. Rates, pathways and drivers for peatland development in the Hudson Bay Lowlands, northern Ontario, Canada. *J. Ecol.* 92: 1036–1053.
Gore, A. J. P. 1983. Introduction. In A. J. P. Gore (Ed.) *Mires: Swamp, Bog, Fen and Moor. Ecosystems of the World, 4A.* Elsevier, New York, NY, pp. 1–34.
Gorham, E. 1991. Northern peatlands: Role in the carbon cycle and probable responses to climatic warming. *Ecol. Appl.* 1: 182–195.
Gorham, E. and J. A. Janssens. 1992. Concepts of fen and bog re-examined in relation to bryophyte cover and the acidity of surface waters. *Acta Soc. Bot. Pol.* 61: 7–20.
Grootjans, A. P., R. van Diggelen, M. J. Wassen, and W. A. Wiersinga. 1988. The effects of drainage on groundwater quality and plant species distribution in stream valley meadows. *Vegetatio* 75: 37–48.
Gumbricht, T. 2012. *Mapping Global Tropical Wetlands from Earth Observing Satellite Imagery.* Working Paper No. 103. CIFOR, Bogor, Indonesia.
Heinselman, M. L. 1963. Forest sites, bog processes, and peatland types in the Glacial Lake Agassiz Region, Minnesota. *Ecol. Monogr.* 33: 327–374.
Heinselman, M. L. 1970. Landscape evolution, peatland types, and the environment in the Lake Agassiz Peatland Natural Area, Minnesota. *Ecol. Monogr.* 40: 235–261.
Hill, B. H., C. M. Elonen, T. M. Jicha, R. K. Kolka, L. L. P. Lehto, S. D. Sebestyen, and L. R. Seifert-Monson. 2014. Ecoenzymatic stoichiometry and microbial processing of organic matter in northern bogs and fens reveals a common P limitation among peatland types. *Biogeochemistry.* 120: 203–224.
Holden, J. and T. P. Burt. 2003. Hydraulic conductivity in upland blanket peat: Measurement and variability. *Hydrol. Process.* 17: 1227–1237.
Ingram, H. A. P. 1982. Size and shape in raised mire ecosystems: A geophysical model. *Nature* 297: 300–303.
Ingram, H. A. P. 1983. Hydrology. In A. J. P. Gore (Ed.) *Mires: Swamp, Bog, Fen and Moor. A. General Studies.* Elsevier Scientific Publishing Company, Amsterdam, The Netherlands, pp. 67–158.
Inisheva, L. I. 2006. Peat soils: Genesis and classification. *Eur. Soil Sci.* 39(7): 699–704.
IPCC. 2006. H. S. Eggleston, L. Buendia, K. Miwa, T. Ngara, and K. Tanabe (Eds.) *2006 IPCC Guidelines for National Greenhouse Gas Inventories, Prepared by the National Greenhouse Gas Inventories Programme*. IGES, Japan.
IPCC. 2014. *2013 Supplement to the 2006 IPCC Guidelines for National Greenhouse Gas Inventories: Wetlands.* IGES, Japan.
IUSS Working Group WRB. 2006. *World Reference Base for Soil Resources*. 2nd ed. World Soil Resources Reports No. 103. FAO, Rome.
Joosten, H. 2009. *The Global Peatland CO_2* Picture: Peatland Status and Drainage Related Emissions in All Countries of the World. Wetlands International, Ede, The Netherlands.
Kane, D. L. and L. D. Hinzman. 1988. Permafrost hydrology of a small arctic watershed. In *Proceedings of the 5th International Conference on Permafrost,* Tapir Publishers, Trondheim, Norway.
Kellogg, L. E. and S. D. Bridgham. 2003. Phosphorus retention and movement across an ombrotrophic–minerotrophic peatland gradient. *Biogeochemistry* 63: 299–315.
Koch, M. S. and K. R. Reddy. 1992. Distribution of soil and plant nutrients along a trophic gradient in the Florida Everglades. *Soil Sci. Soc. Am. J.* 56: 1492–1499.
Koerselman, W., M. B. Van Kerkhoven, and J. T. A. Verhoeven. 1993. Release of inorganic N, P and K in peat soils; effect of temperature, water chemistry and water level. *Biogeochemistry* 20: 63–81.
Kolka, R. K., M. C. Rabenhorst, and D. Swanson. 2011. Histosols. Chapter 33.2. In P. M. Huang, Y. Li, and M. E. Sumner (Eds.) *Handbook of Soil Sciences Properties and Processes*, 2nd ed. CRC Press, Boca Raton, FL. pp. 33.8–33.29.

Korpijaako, M. and N. W. Radforth. 1972. Studies of the hydraulic conductivity of peat. In *Proc. 4th Int. Peat Congr. V. 3*. Otaniemi, Finland, pp. 323–334.

Lee, J. A., M. C. Press, S. Woodin, and P. Ferguson. 1987. Responses to acidic deposition in ombrotrophic mires in the U. K. In T. C. Hutchinson and K. M. Meema (Eds.) *Effects of Atmospheric Pollutants on Forests, Wetlands and Agricultural Ecosystems.* Springer-Verlag, Berlin, Germany, pp. 549–560.

Lehner, B. and P. Döll. 2004. Development and validation of a global database of lakes, reservoirs and wetlands. *J. Hydrol.* 296: 1–22.

Lévesque, M. P. and S. P. Mathur. 1979. A comparison of various means of measuring the degree of decomposition of virgin peat materials in the context of their relative biodegradability. *Can. J. Soil Sci.* 59: 397–400.

Malterer, T. J., E. S. Verry, and J. Erjavec. 1992. Fiber content and degree of decomposition in peats: A review of national methods. *Soil Sci. Soc. Am. J.* 56: 1200–1211.

Mathur, S. P. and R. S. Farnham. 1985. Geochemistry of humic substances in natural and cultivated peatlands. In G. R. Aiken, D. M. McKnight, R. L. Wershaw, and P. MacCarthy (Eds.) *Humic Substances in Soil, Sediment, and Water: Geochemistry, Isolation, and Characterization.* John Wiley & Sons, New York, NY, pp. 53–85.

Melton, J. R., R. Wania, E. Hodson, B. Poulter, B. Ringeval, R. Spahni, T. Bohn. et al. 2013. Present state of global wetland extent and wetland methane modelling: Conclusions from a model intercomparison project (WETCHIMP). *Biogeosciences* 10: 753–788.

Michaelson, G. J., C. L. Ping, and J. M. Kimble. 1996. Carbon storage and distribution in tundra soils of Arctic Alaska U.S.A. *Arct. Alp. Res.* 28: 414–424.

Moore, P. D. and D. J. Bellamy. 1974. *Peatlands*. Springer-Verlag, New York, NY.

Moore, T. R. and R. Knowles. 1989. The influence of water table levels on methane and carbon dioxide emissions from peatland soils. *Can. J. Soil Sci.* 69: 33–38.

National Wetlands Working Group. 1988. *Wetlands of Canada.* Sustainable Development Branch, Environment Canada, Ontario, and Polyscience Publications, Montreal, Quebec, Canada.

National Wetlands Working Group. 1997. *The Canadian Wetland Classification System*. Wetlands Research Centre, University of Waterloo, Waterloo, Ontario, Canada.

Nykanen, H., J. Alm, K. Lang, J. Silvola, and P. J. Martikainen. 1995. Emissions of CH_4, N_2O and CO_2 from a virgin fen and a fen drained for grassland in Finland. *J. Biogeogr.* 22: 351–357.

Ohlson, M. and R. H. Okland. 1998. Spatial variation in rates of carbon and nitrogen accumulation in a boreal bog. *Ecology* 79: 2745–2758.

Ovenden, L. 1990. Peat accumulation in northern wetlands. *Quat. Res.* 33: 377–386.

Page, S. E., J. O. Rieley, and C. J. Banks. 2011. Global and regional importance of the tropical peatland carbon pool. *Glob. Change Biol.* 17: 798–818.

Päivänen, J. 1973. Hydraulic conductivity and water retention in peat soils. *Acta For. Fenn.* 129: 1–70.

Parent, L. E. and J. Caron. 1993. Physical properties of organic soils. In M. R. Carter (Ed.) *Soil Sampling and Methods of Analysis*. CRC Press, Boca Raton, FL, pp. 441–458.

Péwé, T. 1975. *Quaternary Geology of Alaska. Geological Survey Professional Paper 835, U.S. Geological Survey*. U.S. Govt. Printing Office. Washington, DC.

Ping, C. L., J. G. Bockheim, J. M. Kimble, G. J. Michaelson, D. K. Swanson, and D. A. Walker. 1998. Characteristics of cryogenic soils along a latitudinal transect in arctic Alaska. *J. Geophy. Res.* 103: 28917–28928.

Ping, C. L., G. J. Michaelson, and J. M. Kimble. 1997a. Carbon storage along a latitudinal transect in Alaska. *Nutr. Cycl. Agroecosyst.* 49: 235–242.

Ping, C. L., G. J. Michaelson, W. M. Loya, R. J. Candler, and R. L. Malcolm. 1997b. Characteristics of soil organic matter in Arctic ecosystems of Alaska. In R. Lal, J. M. Kimble, and B. A. Stewart (Eds.) *Carbon Sequestration—Advances in Soil Science*. CRC Press, Boca Raton, FL, pp. 157–167.

Racine, C. H. and J. C. Walters. 1994. Groundwater-discharge fens in the Tanana Lowlands, interior Alaska, U.S.A. *Arc. Alp. Res.* 26: 418–426.

Reeve, A. S., R. Evensen, P. H. Glaser, D. I. Siegel, and D. Rosenberry. 2006. Flow path oscillations in transient groundwater simulations of large peatland systems. *J. Hydrol.* 316: 313–324.

Reeve, A. S., D. I. Siegel, and P. H. Glaser. 2000. Simulating vertical flow in large peatlands. *J. Hydrol.* 227: 207–217.
Rieger, S. 1983. *The Genesis and Classification of Cold Soils*. Academic Press, New York, NY.
Riehle, J. R. 1985. A reconnaissance of the major Holocene tephra deposits in the upper Cook Inlet region, Alaska. *J. Volcano Geotherm. Res.* 26: 37–74.
Rydin, H. 1985. Effect of water level on desiccation of *Sphagnum* in relation to surrounding *Sphagna*. *Oikos* 45: 374–379.
Siegel, D. I. 1988. Evaluating cumulative effects of disturbance on the hydrologic function of bogs, fens, and mires. *Environ. Manag.* 12: 621–626.
Siegel, D. I. 1992. Groundwater hydrology. In H. E. Wright Jr., B. A. Coffin, and N. E. Aaseng (Eds.) *The Patterned Peatlands of Minnesota.* University of Minnesota Press, Minneapolis, MN, pp. 163–172.
Siegel, D. I. and P. H. Glaser. 1987. Groundwater flow in a bog/fen complex, Lost River Peatland, northern Minnesota. *J. Ecol.* 75: 743–754.
Silins, U. and R. L. Rothwell. 1998. Forest peatland drainage and subsidence affect soil water retention and transport properties in an Alberta peatland. *Soil Sci. Soc. Am. J.* 62: 1048–1056.
Sjörs, H. 1950. On the relation between vegetation and electrolytes in north Swedish mire waters. *Oikos* 2: 241–257.
Soil Classification Working Group. 1998. *The Canadian System of Soil Classification.* Agriculture and Agri-Food Canada Publication 1646, NRC Research Press, National Research Council of Canada, Ottawa, Canada.
Soil Survey Staff. 1998. *Query for Histosol Soil Components in the National MUIR and STATSGO Data Sets 8/98.* Natural Resource Conservation Service, USDA, Lincoln, NE and Statistical Laboratory, Iowa State University, Ames, IA.
Soil Survey Staff. 2014. *Keys to Soil Taxonomy.* 12th ed. USDA-Natural Resource Conservation Service, Washington, DC.
SSSA. 2008. *Glossary of Soil Science Terms*. Soil Science Society of America, Madison, WI.
Stanek, W. and T. Silc. 1977. Comparisons of four methods for determination of degree of peat humification (decomposition) with emphasis on the von Post method. *Can. J. Soil Sci.* 57: 109–117.
Strack, M, J. M. Waddington, and E. S. Tuittila. 2004. Effects of water table drawdown on northern peatland methane dynamics: Implications for climate change. *Glob. Biogeochem. Cycles* 18: 4003.
Tarnocai, C. 1998. The amount of organic carbon in various soil orders and ecological provinces in Canada. In R. Lal, J. M. Kimble, R. F. Follett, and B. A. Stewart (Eds.) *Soil Processes and the Carbon Cycle*. CRC Press, Boca Raton, FL, pp. 81–92.
Tarnocai, C., J. G. Canadell, E. A. G. Schuur, P. Kuhry, G. Mazhitova, and S. Zimov. 2009, Soil organic carbon pools in the northern circumpolar permafrost region. *Glob. Biogeochem. Cycles* 23: GB2023. doi:10.1029/2008GB003327.
Tarnocai, C., C. A. S. Smith, and C. A. Fox. 1993. *Guidebook—International Tour of Permafrost-Affected Soils in the Yukon and Northwest Territories of Canada*. Centre for Land and Biological Resources, Research Branch, Agriculture Canada, Ottawa, Canada.
Titus, J. E. and D. J. Wagner. 1984. Carbon balance for two *Sphagnum* mosses: Water balance resolves a physiological paradox. *Ecology* 65: 1765–1774.
Updegraff, K., J. Pastor, S. D. Bridgham, and C. A. Johnston. 1995. Environmental and substrate controls over carbon and nitrogen mineralization in northern wetlands. *Ecol. Appl.* 5: 151–163.
Van Cleve, K., F. S. Chapin III, C. T. Dyrness, and L. A. Viereck. 1991. Element cycling in taiga forests: Statefactor control. *BioScience* 41: 78–88.
Verhoeven, J. T. A., E. Maltby, and M. B. Schmitz. 1990. Nitrogen and phosphorus mineralization in fens and bogs. *J. Ecol.* 78: 713–726.
Verry, E. S. 1997. Hydrological processes of natural, northern forested wetlands. In C. C. Trettin, M. F. Jurgensen, D. F. Grigal, M. R. Gale, and J. K. Jeglum (Eds.) *Northern Forested Wetlands: Ecology and Management*. CRC Press, Chelsea, MI, pp. 163–188.
Vitt, D. H. 2006. Functional characteristics and indicators of boreal peatlands. In R. K. Wieder and D. H. Vitt. (Eds.) *Boreal Peatland Ecosystems.* Springer, New York, NY, pp. 9–24.

Vitt, D. H. and W.-L. Chee. 1990. The relationships of vegetation to surface water chemistry and peat chemistry in fens of Alberta, Canada. *Vegetatio* 89: 87–106.
Walbridge, M. R. 1991. Phosphorus availability in acid organic soils of the Lower North Carolina Coastal Plain. *Ecology* 72: 2083–2109.
Waughman, G. J. 1980. Chemical aspects of the ecology of some south German peatlands. *J. Ecol.* 68: 1025–1046.
Weiss, R., J. Alm, R. Laiho, and J. Laine. 1998. Modeling moisture retention in peat soils. *Soil Sci. Soc. Am. J.* 62: 305–313.
Weltzin, J. F., C. Harth, S. D. Bridgham, J. Pastor, and M. Vonderharr. 2001. Production and microtopography of bog bryophytes: Response to warming and water-table manipulations. *Oecologia* 128: 557–565.
Wilkinson, L., M. Hill, and E. Vang. 1992. *Systat: Statistics*. Systat, Inc., Evanston, IL.
Yu, Z., J. Loisel, D. P. Brosseau, D. W. Beilman, and S. J. Hunt. 2010. Global peatland dynamics since the Last Glacial Maximum. *Geophys. Res. Lett.* 37: L13402.
Zoltai, S. C. and C. Tarnocai. 1971. Properties of a wooded palsa in northern Manitoba. *Arc. Alp. Res.* 3: 115–129.

11

Hydric Soil Indicators in Mollisol Landscapes

James A. Thompson and Jay C. Bell

CONTENTS

Introduction

Mollisols are mineral soils that usually develop under prairie vegetation. They are characterized by relatively thick, dark surface horizons resulting from an increased organic matter content (McDaniel et al. 2011), which can present problems for hydric soil identification due to the lack of visible iron-based redoximorphic features (Chapters 7 and 8) in the upper part of the soil profile. This chapter discusses some of the potential problems encountered when delineating hydric soils in Mollisol landscapes and describes specific hydric soil indicators developed for use in delineating hydric Mollisols.

The thick accumulations of organic matter associated with Mollisols are primarily due to the prairie grass vegetation, which has a dense fibrous root system. The roots, which proliferate in the soil even to depths of 75 cm or greater, have a high rate of annual turnover (Dahlman and Kucera 1965). Because a significant portion of the vegetation biomass is within the soil, root exudates and root turnover readily contribute substantial organic matter to the upper portions of these soils. When the grassland vegetation is disturbed by grazing or fire, the copious, fibrous roots of prairie grasses and the roots of leguminous forbs create abundant "ligno-protein" molecules of soil organic matter that resist oxidation and solution (Hole and Nielsen 1970). The presence of abundant Ca stabilizes organic matter and darkens the soil, creating the characteristic deep black soil characteristic of grassland soils (Mollisols) in temperate regions worldwide. Mollisols, however, can also form under forest vegetation. These soils usually are associated with wetter soil environments or high Ca environments in which the organic matter can be both incorporated and stabilized at rates in excess of decomposition.

A Mollisol must have a mollic epipedon, which, by definition, is a thick (≥25 cm), dark (moist Munsell color of value ≤3 and chroma ≤3), well-structured surface layer that has high organic carbon content (≥6 g kg^{-1}), and has a base saturation ≥50% throughout (Soil Survey Staff 1999). Mollisols constitute approximately 22% of the total land area in the United States (Brady and Weil 1999) and are commonly found throughout the upper Midwest through the Central Plains, in association with historic grassland ecosystems (Pieper 2005). Additional areas of Mollisols in the United States are found in the Palouse area of Washington, Oregon, and Idaho (Figure 11.1). Globally, Mollisols are primarily associated with subhumid-to-semiarid climates where grassland vegetation and calcareous parent materials promote the accumulation of organic matter in the upper horizons (Soil Survey Staff 1999; McDaniel et al. 2011), including eastern Europe; Asia, from Turkey and the Ukraine eastward across Russia; the pampas region of South America; parts of Mexico and Central America (Liu et al. 2012); and under pergelic soil temperature regime in arctic Alaska (Ping and Michaelson 2014).

Aquolls and Albolls are Mollisols that formed in seasonally saturated soil conditions. By definition, Aquolls and Albolls are Mollisols with an aquic moisture regime and one or more of several diagnostic soil horizons. These include a histic epipedon above the mollic, redox concentrations within the mollic, or a gleyed subsurface horizon directly below the mollic epipedon or within 75 cm of the mineral soil surface (Soil Survey Staff 1999). In the United

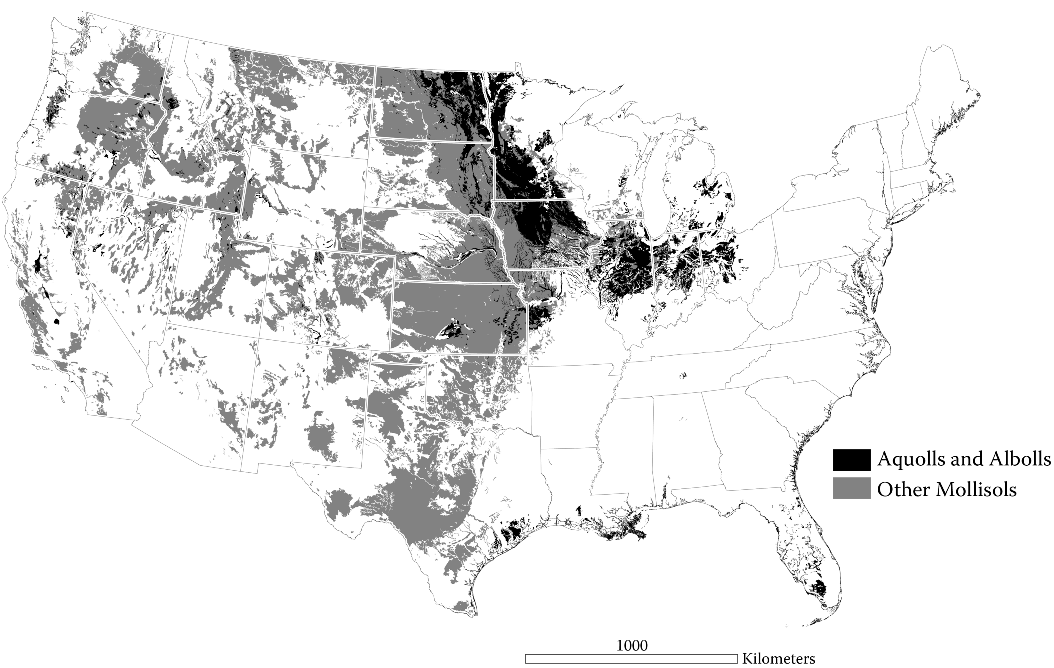


FIGURE 11.1
Generalized map showing the locations of major land areas of Mollisols in the conterminous United States derived from the STATSGO2 database (Soil Survey Staff 2006). Shaded areas represent landscapes where the dominant soil order is Mollisols; dark areas are where wet Mollisols (Aquolls and Albolls) represent 25% or more of the landscape (S. W. Waltman, 2015, personal communication). Smaller areas of Mollisols, including Aquolls and Albolls, are found where vegetation, parent materials, and/or local hydrology favor organic matter accumulation.

States, extensive areas of Aquolls and Albolls are found in broad, flat landscapes with poor natural drainage, such as the Red River Valley of Minnesota and North Dakota, flood plains of the Mississippi, Missouri, Ohio, and Wabash Rivers, and the coastal plains of Louisiana and Texas. These are potentially hydric Mollisols that commonly occur in local depressions where rainfall and slope water can accumulate or in areas of groundwater discharge.

In Mollisol landscapes, delineators of wetlands have had difficulty in separating hydric and nonhydric soils. Common problems are the masking of visible morphologic indicators of hydric soil conditions by the abundant soil organic matter, and the presence of gray-colored carbonates that may mimic accepted hydric soil indicators. Because of these special problems encountered in Mollisols, the following section focuses on hydric soil indicators. In particular, the focus will be on Aquolls in the humid region of the Prairie Pothole glaciated area of the United States.

Soil-Forming Processes in Wet Mollisols

Organic Matter Dynamics

There are several factors that favor the development of Mollisols, including semiarid-to-subhumid climates, grassland vegetation, and calcium-rich parent materials. These factors control the amount of organic materials added to the soil by favoring (i) increased belowground biomass production, (ii) deposition of lignin-rich residues, and (iii) development of stabilizing bonds with Ca that slow the rates of organic matter decomposition (Buol et al. 2011). The dense, fibrous root systems of prairie grasses with substantial annual root turnover and high lignin contents of the residue promote high ($\geq$6 g kg^{-1}) organic matter levels to depths that are greater in Mollisols than in soils of other orders (Figure 11.2). The extensive root system also favors efficient nutrient cycling within the upper part of the soil, which prevents the loss of organic and mineral materials from below the root zone. A combination of the chemical composition of prairie grasses and the calcareous nature of many prairie soils leads to the formation of stable Ca-organic and clay-organic complexes in the upper part of the soil profile.

Increased organic matter additions and reduced organic matter losses result in higher organic matter contents in wetter Mollisols. For example, along a soil moisture gradient in the northern Great Plains, Munn et al. (1978) found that annual plant productivity increases as soil moisture increases. This results in greater organic matter accumulation in the soil. Also, in wetter soil environments, the rates of organic matter decomposition are substantially lowered, mainly due to the lack of oxygen in the soil. Anaerobic decomposition of soil organic matter is slower than aerobic decomposition because anaerobic conditions produce end products that inhibit microbial activity or are toxic to soil microorganisms (Ross 1989). Increased organic matter incorporation coupled with decreases in soil organic matter losses due to wetness results in wet Mollisols that have even thicker and darker surface horizons, such as the soils examined by Richardson and Bigler (1984), Thompson and Bell (1996, 1998), and Reuter and Bell (2003).

Mollisol Landscapes

The morphological differences in Mollisols that result from increased wetness can most easily be seen along soil moisture gradients found along many hillslopes (Figure 11.3).

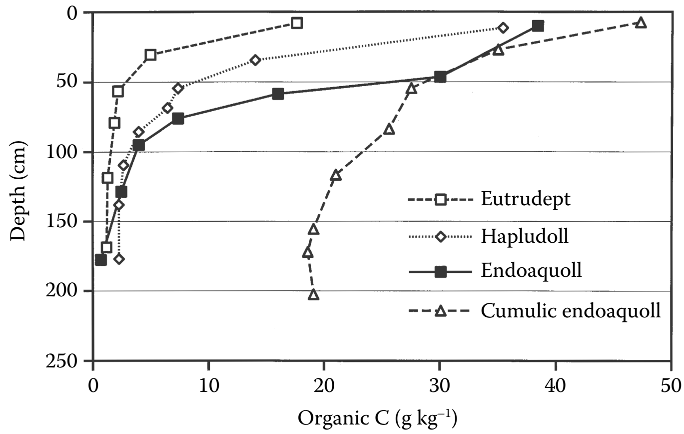


FIGURE 11.2
Soil organic carbon (C) with depth for an Inceptisol and three Mollisols of varying degree of wetness and mollic development. Differences among the Eutrudept, the Hapludoll, and the Endoaquoll are greatest in the upper 50 cm; organic C contents are similar below 1 m. The Cumulic Endoaquoll differs in that the organic C content remains high to a depth below 2 m. (Data derived from Thompson, J. A. and J. C. Bell. 1996. *Soil Sci. Soc. Am. J.* 60: 1979–1988.)

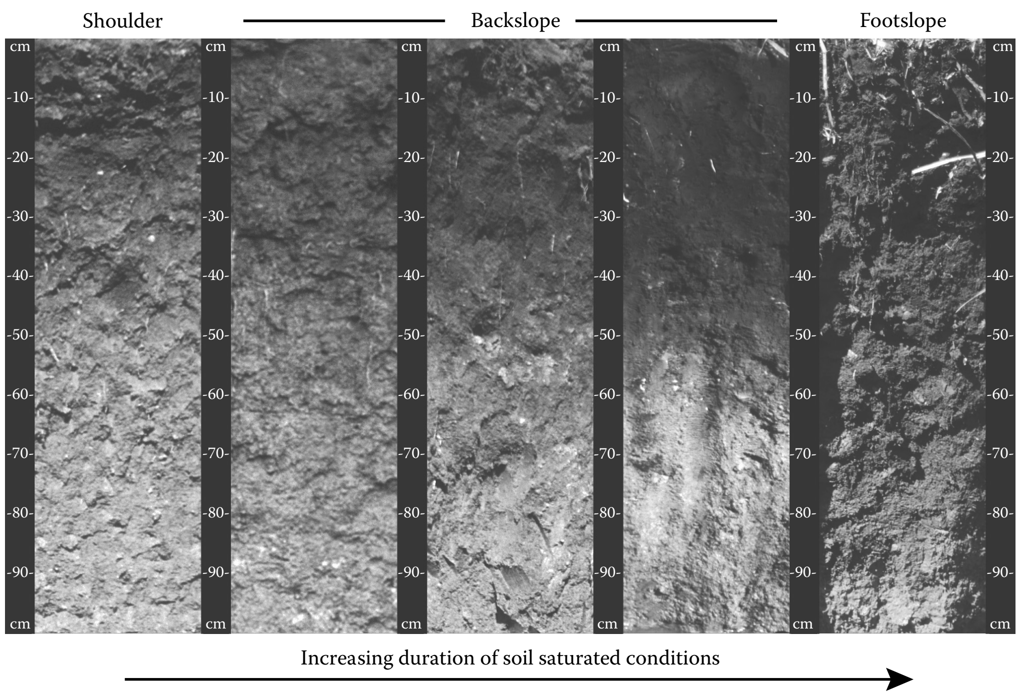


FIGURE 11.3
Soils along a hillslope transect in a Mollisol landscape in southeastern Minnesota. (From Reuter, R. J. and J. C. Bell. 2003. *Soil Sci. Soc. Am. J.* 67(1): 365–372.) Note the thickening and darkening of the surface horizons. Though not visible here, the subsurface horizons become yellower in color as relative wetness of these soils increases. (Photos by J. A. Thompson.)

In general, with increasing soil wetness, the A horizon(s) will (i) increase in thickness, (ii) decrease in Munsell value, and (iii) decrease in Munsell chroma. The wettest soils within these landscapes can have 1–2 m of black (N 2/0) soil at the surface. These trends in soil morphology from well-drained to very poorly drained soils in Mollisol hillslopes have been described in Minnesota (Bell et al. 1995, 1996; Thompson and Bell 1996, 1998; Bell and Richardson 1997; Reuter and Bell 2003), Iowa (James and Fenton 1993; Khan and Fenton 1994, 1996), elsewhere in the Prairie Pothole Region (Richardson et al. 1994), and other Mollisol landscapes (Abtahi and Khormali 2001).

The color of the soil directly below the mollic epipedon may also be a useful indicator of wetness in Mollisol landscapes. In general, a low-chroma (≤2) matrix is indicative of prolonged saturated soil conditions (Vepraskas 1994). However, in areas where parent materials are naturally gray, observation of low-chroma colors alone does not confirm the presence of seasonally saturated soil conditions. In landscapes underlain by low-chroma parent materials, it is advisable to look at relative differences in the soil color from well-drained soils to very poorly drained soils instead of relying on Munsell value and chroma observations from individual points. With increasing soil wetness, there is a decrease in the chroma and an increase in the value of the upper B horizon. Thompson and Bell (1996, 1998) reported changes in subsoil color of only 1 or 2 chroma units. More pronounced dulling of the subsoil matrix color is seen in data presented by Khan and Fenton (1994, 1996) and Reuter and Bell (2003). Their data show a decrease in subsoil chroma from 4 to 2, an increase in value from 5 to 6, and a more yellow hue, with a change from 10YR to 5Y.

Redoximorphic features that indicate soil wetness are sometimes observed in or directly below the A horizon(s). In general, the depth to high-chroma mottles is shallower in wetter Mollisols. However, the lack of observable iron-based redoximorphic features does not preclude the occurrence of prolonged soil wetness in the mollic epipedon (Reuter and Bell 2003). The lack of observable redoximorphic features has been attributed to masking by organic coatings on ped and particle surfaces (Parker et al. 1985) that are frequently a direct result of the anaerobic conditions that inhibit organic matter decomposition. Others have attributed the lack of observable redoximorphic features in mineral soils with higher organic matter contents to the formation of Fe-organic complexes that prevent the formation of iron oxides (Schwertmann et al. 1986; Tan 1986; Wheeler et al. 1999).

Hillslope Processes

Erosion, groundwater discharge, and evaporative discharge on depression edges can also create problems for hydric soil delineation. Because many wetter Mollisols are found in local landscape positions that collect slope water such as depressions and flood plains (Richardson et al. 1994; Thompson and Bell 1998; Bedard-Haughn and Pennock 2002; Reuter and Bell 2003; Debelis et al. 2005), they collect any materials carried by the water, for example, sediments or dissolved solids (Hayashi et al. 1998; Knox 2006). Erosion of A horizon material from upslope positions with redeposition in lower landscape positions can add significant amounts of darker soil materials to the surface of soils in lower landscape positions. Many wet Mollisols found in lower hillslope positions have buried A horizons. However, the more recently deposited surface materials in these wetter Mollisols tend to be lighter in color (chroma 1 or 2) than the underlying buried A horizon, which tends to be black (N 2/0).

Along with accumulation of organic matter, the depth to carbonates in Mollisols can impart information on hillslope hydrology (Chapter 3). The subhumid-to-semiarid

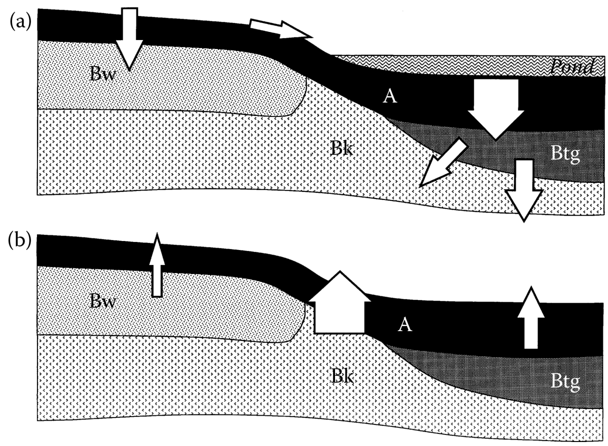


FIGURE 11.4
Generalized local hydrology of a wetland basin showing seasonal shifts between depression-focused recharge and edge-focused discharge. (a) During the wet season, water flow into the depression and infiltration promotes saturated downward flow. (b) During the dry season, evapotranspiration promotes unsaturated upward flow. The edges of the depression have the longest period of time with upward flow and lack downward flow in the wet periods, producing shallow calcic horizons. Arrows are proportional to the amount of water flow. (Figure by J. L. Richardson, personal communication and used by permission.)

climates in the upper Midwest favor retention of carbonates that are derived from calcareous parent materials. The depth to these carbonates, especially in the wetland areas, is related to wetland hydrology. In groundwater recharge wetlands where water flow is predominantly downward, leaching of carbonates can produce soil profiles that are relatively free of carbonates in the upper 1–2 m (King et al. 1983; Knuteson et al. 1989; Mausbach and Richardson 1994; Richardson et al. 1994; Abtahi and Khormali 2001; Bedard-Haughn and Pennock 2002). In groundwater discharge wetlands, where water flow is predominantly upward, carbonate contents are high throughout the profile, sometimes even accumulating at the soil surface (Arndt and Richardson 1988; Khan and Fenton 1994; Richardson et al. 1994; Bedard-Haughn and Pennock 2002).

The development of a highly calcareous soil horizon at the edge of depressions that trap surface water is a common feature in most young glaciated landscapes (Richardson et al. 1994; Chapter 3) extending from humid climates in Iowa (Steinwand and Fenton 1995), northward into subhumid areas of North Dakota (Steinwand and Richardson 1989), and into the semiarid Canadian prairies (Miller et al. 1985) (Figure 11.4). The edges of the wetland have plants and a near-surface water table that combine to evaporate and transpire far more water from the soil than that which moves downward into the soil. The result is a strong reversal of leaching in which dissolved solids are transferred from the landscape and the wetland to the edge and concentrated by evapotranspiration. The concentrations reach levels that allow for the formation of calcite and sometimes gypsum. These evaporites are naturally a gray color and may resemble depletions. Their occurrence can create a false indication of redoximorphic features. However, depletions on surfaces can be noted in these gray evaporites. If the iron-depleted surfaces in gray calcareous horizons exceed 5% and occur within 30 cm of the surface, soil scientists in North Dakota use these features as a positive identification of a hydric soil, verifying the presence of a depletion feature (J. L. Richardson, personal communication).

Hydric Soil Indicators in Wet Mollisols

Hydric Soil Definition and Criteria

By definition, a hydric soil is a soil formed under saturated, flooded, or ponded conditions long enough during the growing season to develop anaerobic conditions in the upper part (Federal Register 1994; Chapter 2). In most soils, the result of anaerobic conditions is reflected in the general hydric soil indicators (Environmental Laboratory 1987; Chapters 8 and 9): organic soil or histic epipedon; sulfidic material; gleyed, low chroma, and low chroma/mottled soils; or iron and manganese concretions. However, some soils have morphologies that are difficult to identify as hydric because of (i) low chroma or red parent materials, (ii) high or low organic matter contents, (iii) high pH (which inhibits iron reduction), or (iv) natural or anthropogenic site disturbance. Hydric Mollisols are particularly difficult to identify because of high organic matter content and natural mixing by soil organisms. Consequently, Mollisols were designated as problem soils that require special consideration for the development of reliable field indicators of hydric soil conditions (Federal Interagency Committee for Wetlands Delineation 1989).

Field Indicators of Hydric Soils for Mollisols

Among the soil morphological features used to identify hydric soils in the *Field Indicators of Hydric Soils in the United States* (USDA-NRCS 2010), six field indicators are targeted for soils with thick, dark A horizons. All indicators are the result of field-based investigations of actual wetlands and associated soils, and are intended for delineation of the edge of wetlands, and not the wetter interiors. The indicators reflect the available information derived from hydrology, vegetation, landscape position, and the best professional judgment of wetland scientists. The indicators are officially accepted by the National Technical Committee for Hydric Soils and have been widely tested.

The field indicators developed for use in Mollisol landscapes reflect the thickening and darkening of the surface horizons and the dulling and gleying of the subsurface horizons associated with increasing wetness. Soil morphologies that indicate hydric soil conditions in soils with dark surface horizons include: (i) a depleted matrix immediately below a dark surface that does not have observable redoximorphic features (indicators A11 and A12, Chapter 9), (ii) a dark surface with redoximorphic concentrations or depletions (indicator F6), (iii) sandy soils with low-chroma colors directly below a dark surface (indicators S7, S8, A11, and A12), or (iv) soils in concave landforms with low-chroma colors directly below a dark surface (indicator F13) (USDA-NRCS 2010). While the indicators will work in most field situations, there are instances where special care must be taken to properly identify hydric Mollisols. As is noted in the field indicators' guide, a soil without an indicator may still be classified as hydric.

An important provision in the field indicators' guide is included for soils where anaerobic conditions develop within the upper 30 cm, but short durations of saturated conditions in the upper part are not sufficient to lead to the development of anaerobic conditions that result in low-chroma soil colors throughout the upper 30 cm (USDA-NRCS 2010):

> All mineral layers above any layers meeting the requirements of any indicators…have dominant chroma of 2 or less, or the thickness of the layer(s) with dominant chroma of more than 2 is less than 15 cm.

In many soil landscapes we have observed, this exception can be applied to profiles with surface erosional sediments that presumably were deposited following European settlement after hydric soil morphology developed. The lighter-colored layers or horizons targeted by this provision must be less than 15 cm thick.

We have observed soils from lower landscape positions in steeply sloping agricultural sites in west-central Minnesota that have up to 90 cm of lighter-colored (10YR 3/2 and 2/2) materials deposited above >60 cm of black (N 2/0) soil with distinct (7.5YR 5/8) redox concentrations (unpublished data). These soils may have formed under hydric soil conditions because of the thick, black buried A horizon and the toeslope landscape position. However, these soils would not be considered hydric based on the indicators due to the thick overlying accumulation of lighter-colored erosional sediments. While this represents an extreme example, any accumulation of erosional sediments greater than 15 cm would not permit a soil to be classified as hydric based on the current field indicators. In steeply sloping and/or intensively farmed landscapes, the presence of excessively thick erosional sediments is common.

A potentially useful iron-based field indicator of hydric soils in Mollisols is the presence of oxidized root channels (indicator F6, redox dark surface) (Mendelssohn 1993; Mendelssohn et al. 1995). When present, these features are prominent against the dark matrix colors of the mollic epipedon. We have observed oxidized root channels in hydric Mollisols in the late spring. However, they were not present in the same soils in the late summer of the previous year. Oxidized root channels may be ephemeral features in these soils and, therefore, only indicate recent soil anaerobic conditions. Also, cultivation or other soil mixing can obliterate these features.

Landscape Position

The importance of landscape position is recognized in certain field indicators. Our experience is that in Mollisol landscapes, landscape position is particularly important for hydric soil determinations. In the Prairie Pothole Region, water accumulates in closed or nearly closed depressions. These depressions focus water from the local landscape. Hydric soils may occur in the depressions, and the nearly level toeslopes to concave footslopes surrounding the depression (Mausbach and Richardson 1994; Richardson et al. 1994; Thompson et al. 1997). In areas with open drainage, hydric soils may develop in flood plains, concave and convergent slopes, and areas with nearly level toeslopes to concave footslopes below steep slopes (Mausbach and Richardson 1994; Thompson et al. 1998). In general, converging slopes tend to concentrate water and allow it to accumulate long enough for hydric soils to develop (Mausbach and Richardson 1994; Chapter 3). However, because other processes, such as erosion and deposition, contribute to the development of thick, dark surface horizons, some cumulic soils in the low-sloping landscape position may not be hydric. Thompson and Bell (1996) describe a soil with over 1 m of black (10YR 2/1) surface horizons, but their data indicate that the water table was not within 80 cm of the soil surface during 2 years of monitoring.

Differences in soil hydrology and the resultant soil morphology between soils of different landscape positions on a single hillslope are illustrated in Figure 11.5. A very poorly drained soil located in a drainageway of a low-order, intermittent stream (Figure 11.5a) shows high water tables throughout the year. The thick, black (N 2/0) surface horizons and high organic matter content reflect the high water table conditions observed in this soil. In a well-drained soil located on the summit position of this same landscape (Figure 11.5b), water tables are lower and fluctuations are greater than in the very poorly drained soil.

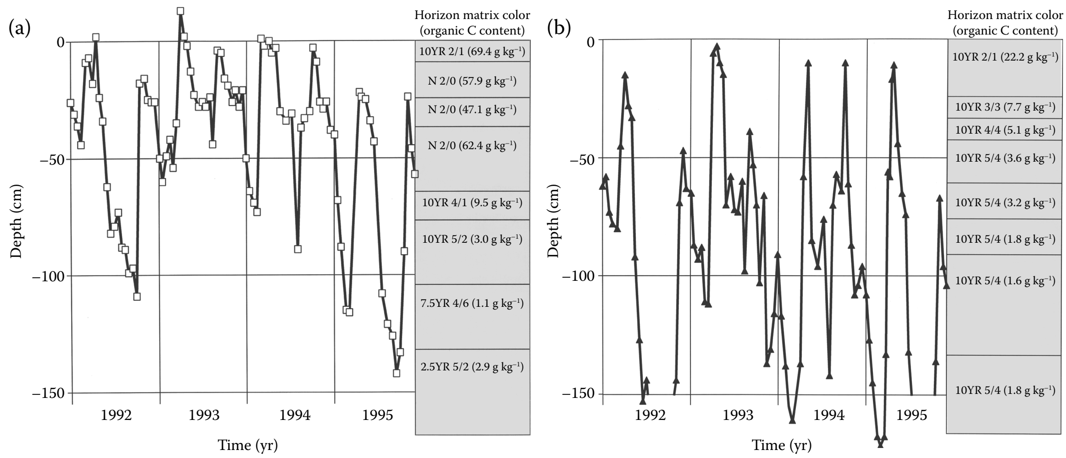


FIGURE 11.5
Observed water table position and morphological and chemical properties of (a) a very poorly drained soil in a drainageway landscape position, and (b) a well-drained soil in a summit landscape position. The water table is above or near the soil surface in the spring and fall at both landscape positions. However, while the water table falls considerably in the well-drained soil, in the very poorly drained soil, it remains within 50 cm of the soil surface for extended periods even during the summer. These differences in hydrology are reflected in the higher organic C content and thicker, darker soil colors of the wetter soil. (Data derived from Thompson, J. A. and J. C. Bell. 1998. *Soil Sci. Soc. Am. J.* 62: 1116–1125.)

While this soil is still a Mollisol, the thickness, darkness, and organic matter contents of the upper horizons are considerably less than the very poorly drained soil.

Profile Darkness Index

Thompson and Bell (1996) proposed a soil color index, the profile darkness index (PDI), that quantified the trends of increasing A horizon thickness and darkness in Mollisol catenas. Calculated for each horizon with a Munsell value ≤3 and a Munsell chroma ≤3, PDI is equal to the sum (over all dark horizons within the profile) of the horizon thickness, divided by the quantity one plus the Munsell value times the Munsell chroma:

$$\text{PDI} = \sum_{i=1}^{n} \frac{\text{A-horizon thickness}_i}{(V_i C_i + 1)}$$

where thickness is measured in centimeters, V is Munsell value, C is Munsell chroma, and n is the total number of A horizons described. A plot of PDI along a transect from summit to depression in a Mollisol landscape in west-central Minnesota (Figure 11.6) illustrates the landscape-scale trends in PDI, which reflect the observed thickening and darkening of the surface horizons as soil wetness increases.

The use of PDI for hydric soil identification and delineation requires setting a threshold value that separates hydric from nonhydric soils. On the basis of variations in PDI among three study sites (Thompson and Bell 1996, 1998), the threshold value would be specific to soils of similar climate and/or parent materials. These differences among only three sites accentuate the necessity for regionalization of this approach. As with other indicators of hydric soils, the PDI threshold value will change among climates and parent materials.

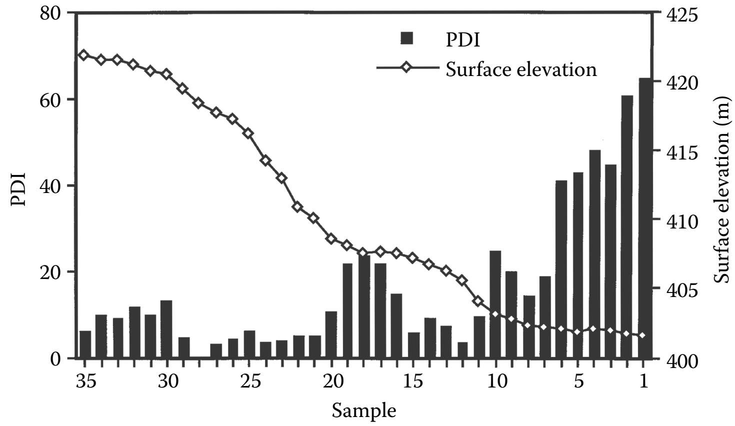


FIGURE 11.6
Variation in the PDI and surface elevation along a hillslope transect in west-central Minnesota. The PDI values are higher where slope gradients are lowest (e.g., from samples 1 to 10, 16 to 19, and 30 to 35), with a general increase across the transect toward the wetter, lower landscape positions (e.g., the toeslope [sample 5] and depression [sample 1]). (Figure modified from Thompson, J. A. and J. C. Bell. 1996. *Soil Sci. Soc. Am. J.* 60: 1979–1988.)

Other authors have since applied PDI in additional settings. For example, Reuter and Bell (2003) found PDI to be strongly correlated with duration of saturation in a Mollisol landscape in southern Minnesota. Conversely, Fiedler and Sommers (2004), working in a non-Mollisol landscape, found no relationships between PDI and soil wetness or redox conditions.

Conclusions

The identification of hydric soils in Mollisol landscapes is problematic because of surface accumulations of thick, dark, and organic-rich soil materials, particularly in wet Mollisols. While common Fe-based field indicators of hydric soils are not always useful for the identification of hydric Mollisols, other diagnostic soil and landscape features can be used. The accumulation of organic C to form thick, dark surface horizons with the presence of redoximorphic features—or without redoximorphic features but with a reduced or depleted horizon immediately below the dark surface—usually reflects seasonal saturated conditions in Mollisol landscapes. In addition to soil morphology, descriptions of landscape position, which aid in understanding where in the landscape water will tend to accumulate, are often also useful in field identification of hydric soils.

References

Abtahi, A. and F. Khormali. 2001. Genesis and morphological characteristics of Mollisols formed in a catena under water table influence in southern Iran. *Comm. Soil Sci. Plant Anal.* 32(9–10): 1643–1658.

Arndt, J. L. and J. L. Richardson. 1988. Hydrology, salinity, and hydric soil development in a North Dakota Prairie-Pothole wetland system. *Wetlands* 8: 94–108.

Bedard-Haughn, A. K. and D. J. Pennock. 2002. Terrain controls on depressional soil distribution in a hummocky morainal landscape. *Geoderma* 110(3): 169–190.

Bell, J. C., C. A. Butler, and J. A. Thompson. 1996. Soil hydrology and morphology of three Mollisol hydrosequences in Minnesota. In J. Wakely, and S. Sprecher (Eds.) *Preliminary Investigations of Hydric Soil Hydrology and Morphology in Alaska, Indiana, North Dakota, Minnesota, and Oregon.* Wetlands Research Program Technical Report. U.S. Army Corps of Engineers, Waterways Experiment Station, Vicksburg, MS.

Bell, J. C. and J. L. Richardson. 1997. Aquic conditions and hydric soil indicators for Aquolls and Albolls. In M. J. Vepraskas, and S. Sprecher (Eds.) *Aquic Conditions and Hydric Soils: The Problem Soils.* SSSA Special Publication of the Soil Science Society of America, Madison, WI, pp. 23–40.

Bell, J. C., J. A. Thompson, and C. A. Butler. 1995. Morphological indicators of seasonally saturated soils for a hydrosequence in Southeastern Minnesota. *J. Minn. Acta Sci.* 59: 25–34.

Brady, N. C. and R. R. Weil. 1999. *The Nature and Properties of Soils*, 12th ed. Prentice-Hall, Inc., New Jersey.

Buol, S. W., R. J. Southard, R. C. Graham, and P. A. McDaniel. 2011. *Soil Genesis and Classification*. John Wiley & Sons.

Dahlman, R. C. and C. L. Kucera. 1965. Root productivity and turnover in native prairie. *Ecology* 46: 84–89.

Debelis, S. P., A. A. Bozzo, M. B. Barrios, and A. Buján. 2005. The relationship between soil characteristics and vegetation as a function of landform position in an area of the Flooding Pampa. *Span. J. Agric. Res.* 3(2): 232–242.

Environmental Laboratory. 1987. *Corps of Engineers Wetlands Delineation Manual*. Technical Report Y-87-1. U.S. Army Engineer Waterways Experiment Station, Vicksburg, MS.

Federal Interagency Committee for Wetlands Delineation. 1989. *Federal Manual for Identifying and Delineating Jurisdictional Wetlands.* U.S. Army Corps of Engineers, U.S. Environmental Protection Agency, U.S. Fish and Wildlife Service, and USDA Soil Conservation Service, Washington, DC. Cooperative technical publication.

Federal Register. 1994. *Definition of Hydric Soils.* USDA-Natural Resources Conservation Service, Vol. 59(133), Wednesday, July 13, U.S. Government Printing Office, Washington, DC, pp. 35681.

Fiedler, S. and M. Sommer. 2004. Water and redox conditions in wetland soils—Their influence on pedogenic oxides and morphology. *Soil Sci. Soc. Am. J.* 68(1): 326–335.

Hayashi, M., G. van der Kamp, and D. L. Rudolph. 1998. Water and solute transfer between a prairie wetland and adjacent uplands, 2. Chloride cycle. *J. Hydrol.* 207(1): 56–67.

Hole, F. D. and G. A. Nielsen. 1970. Some processes of soil genesis under prairie. *Proceedings of a Symposium on Prairie and Prairie Restoration*. Knox College, Galesburg, IL, pp. 28–34.

James, H. R. and T. E. Fenton. 1993. Water tables in paired artificially drained and undrained soil catenas in Iowa. *Soil Sci. Soc. Am. J.* 57: 774–781.

Khan, F. A. and T. E. Fenton. 1994. Saturated zones and soil morphology in a Mollisol catena of central Iowa. *Soil Sci. Soc. Am. J.* 58: 1457–1464.

Khan, F. A. and T. E. Fenton. 1996. Secondary iron and manganese distributions and aquic conditions in a Mollisol catena of central Iowa. *Soil Sci. Soc. Am. J.* 60: 546–551.

King, G. J., D. F. Acton, and R. J. St. Arnaud. 1983. Soil–landscape analysis in relation to soil distribution and mapping at a site within the Wyburn Association. *Can. J. Soil Sci.* 63: 657–670.

Knox, J. C. 2006. Floodplain sedimentation in the Upper Mississippi Valley: Natural versus human accelerated. *Geomorphology* 79(3): 286–310.

Knuteson, J. A., J. L. Richardson, D. D. Patterson, and L. Prunty. 1989. Pedogenic carbonates in a Calciaquoll associated recharge wetland. *Soil Sci. Soc. Am. J.* 53: 495–499.

Liu, X., C. L. Burras, Y. S. Kravchenko, A. Duran, T. Huffman, H. Morras, G. Studdert, X. Zhang, R. M. Cruse, and X. Yuan. 2012. Overview of Mollisols in the world: Distribution, land use and management. *Can. J. Soil Sci.* 92(3): 383–402.

Mausbach, M. J. and J. L. Richardson. 1994. Biogeochemical processes in hydric soil formation. *Curr. Top. Wetland Biogeochem.* 1: 68–127.

McDaniel, P. A., R. G. Darmody, J. C. Bell, and D. G. McGahan. 2011. Mollisols. In P. M. Huang, Y. Li, and M. E. Sumner (Eds.) *Handbook of Soil Science*, 2nd ed. CRC Press, Boca Raton, FL, pp. 33-97–33-113.

Mendelssohn, I. A. 1993. *Factors Controlling the Formation of Oxidized Root Channels: A Review and Annotated Bibliography*. Technical Report WRP-DE-5. U.S. Army Engineer Waterways Experiment Station, Vicksburg, MS.

Mendelssohn, I. A., B. A. Kleiss, and J. S. Wakeley. 1995. Factors controlling the formation of oxidized root channels: A review. *Wetlands* 15: 37–46.

Miller, J. J., D. F. Acton, and R. J. St. Arnaud. 1985. The effect of groundwater on soil formation in a morainal landscape in Saskatchewan. *Can. J. Soil Sci.* 65: 293–307.

Munn, L. C., G. A. Nielsen, and W. F. Mueggler. 1978. Relationships of soils to mountain and foothill range habitat types and production in western Montana. *Soil Sci. Soc. Am. J.* 42: 135–139.

Parker, W. B., S. P. Faulkner, and W. H. Patrick. 1985. Soil wetness and aeration in selected soils with aquic moisture regimes in the Mississippi and Pearl River deltas. In *Wetland Soils: Characterization, Classification, and Utilization*. International Rice Research Institute, Los Baos, Laguna, Philippines, pp. 91–107.

Pieper, R. D. 2005. *Grasslands of Central North America. Grasslands of the World*. Food and Agricultural Organization of the United Nations, Rome, Italy, pp. 221–263.

Ping, C. L. and G. J. Michaelson. 2014. *Guidebook: Alaska Soil Geography, Permafrost-Affected Soils along the Dalton Highway, Alaska*. Bulletin 117, School of Natural Resources and Extension University of Alaska Fairbanks, Alaska.

Reuter, R. J. and J. C. Bell. 2003. Hillslope hydrology and soil morphology for a wetland basin in south-central Minnesota. *Soil Sci. Soc. Am. J.* 67(1): 365–372.

Richardson, J. L., J. L. Arndt, and J. Freeland. 1994. Wetland soils of the Prairie Potholes. *Adv. Agron.* 52: 121–171.

Richardson, J. L. and R. J. Bigler. 1984. Principal component analysis of Prairie Pothole soils in North Dakota. *Soil Sci. Soc. Am. J.* 48: 1350–1355.

Ross, S. 1989. *Soil Processes: A Systematic Approach*. Routledge, New York.

Schwertmann, U., H. Kodama, and W. R. Fischer. 1986. Mutual interactions between organics and iron oxides. In P. M. Huang, and M. Schnitzer (Eds.) *Interactions of Soil Minerals with Natural Organics and Microbes*. Soil Science Society of America, Madison, WI, USA. Special Publication Number 17, pp. 223–250.

Soil Survey Staff. 1999. *Soil Taxonomy: A Basic System of Soil Classification for Making and Interpreting Soil Surveys*. 2nd edition. Natural Resources Conservation Service. U.S. Department of Agriculture Handbook 436.

Soil Survey Staff. 2006. *Digital General Soil Map of the United States (STATSGO2)*. Digital Map and Associated Attribute Tables. USDA Natural Resources Conservation Service Soil Data Mart.

Steinwand, A. L. and T. E. Fenton. 1995. Landscape evolution and shallow groundwater hydrology of a till landscape in central Iowa. *Soil Sci. Soc. Am. J.* 59: 1370–1377.

Steinwand, A. L. and J. L. Richardson. 1989. Gypsum occurrence in soils on the margin of semipermanent Prairie Pothole wetlands. *Soil Sci. Soc. Am. J.* 53: 836–842.

Tan, K. H. 1986. Degradation of soil mineral by organic acids. In P. M. Huang, and M. Schnitzer (Eds.) *Interactions of Soil Minerals with Natural Organics and Microbes*. Soil Science Society of America, Madison, WI, USA. Special Publication Number 17, pp. 1–27.

Thompson, J. A. and J. C. Bell. 1996. Color index for identifying hydric soil conditions in seasonally-saturated Mollisols. *Soil Sci. Soc. Am. J.* 60: 1979–1988.

Thompson, J. A. and J. C. Bell. 1998. Hydric conditions and hydromorphic properties within a Mollisol catena in southeastern Minnesota. *Soil Sci. Soc. Am. J.* 62: 1116–1125.

Thompson, J. A., J. C. Bell, and C. A. Butler. 1997. Quantitative soil–landscape modeling for estimating the areal extent of hydromorphic soils. *Soil Sci. Soc. Am. J.* 61: 971–980.

Thompson, J. A., J. C. Bell, and C. W. Zanner. 1998. Hydrology and hydric soil extent within a Mollisol catena in southeastern Minnesota. *Soil Sci. Soc. Am. J.* 62: 1126–1133.

United States Department of Agriculture, Natural Resources Conservation Service. 2010. Field indicators of hydric soils in the United States, Version 7.0. In L. M. Vasilas, G. W. Hurt, and C. V. Noble (Eds.) *USDA, NRCS, in cooperation with the National Technical Committee for Hydric Soils*.

Vepraskas, M. J. 1994. *Redoximorphic Features for Identifying Aquic Conditions*. Technical Bulletin 301. North Carolina Agricultural Research Service, North Carolina State University, Raleigh, NC.

Wheeler, D. B., J. A. Thompson, and J. C. Bell. 1999. Laboratory comparison of soil redox conditions between red soils and brown soils in Minnesota, USA. *Wetlands* 19(3): 607–616.

12

Hydric Soils and Wetlands in Riverine Systems

Patrick J. Drohan, David L. Lindbo, and Jimmie L. Richardson

CONTENTS

Introduction

The term *riverine system* as used here refers to a river or stream valley measured from the stream channel to the valley edge, including floodplain or terraces that can be inundated or flooded frequently. An active riverine system is one that lacks upstream dams, has not been channelized or has constructed levees, or has been entrenched to a degree that flooding no longer occurs. They occur throughout the world in virtually every climate. In the following section, we concentrate our comments on riverine systems that are related to meandering rivers. Meandering rivers create floodplains that are associated with extensive wetlands. Although rivers themselves do not account for a large percentage of the Earth's surface, their influence is nonetheless of paramount importance. Riverine systems contain some of the world's most fertile agricultural and silvicultural lands, and are adjacent to numerous large cities. As a result, land within these systems is under increasing pressure from development and is exposed to a diversity of environmental hazards.

The riverine system is formed and constantly modified by fluvial (channel stream flow) and other hydrologic processes. These processes influence wetland occurrence and extent.

Because of the fluvial and groundwater interaction with landform, soil, and vegetation, our discussion will start with these dynamic processes. The discussion of features within the riverine system focuses on the floodplain, which is their most common and often-defining feature. Wetlands and hydric soils also occur at the interface of uplands and river valley terraces, at the headwaters of the riverine system as groundwater seeps (see Chapter 3), and they are associated with oxbows and related features.

The Riverine System

We believe that John Playfair (1802) best expressed the idea of streams and their valleys nearly 200 years ago:

> Every river appears to consist of a main trunk, fed from a variety of branches, each running in a valley proportioned to its size, and all of them together form a system of vallies (*sic*), communical with one another, and having such a nice adjustment of their declivities, that none of them join the principal valley, either on too high or too low a level, a circumstance which would be infinitely improbable if each of these vallies were not the work of the stream that flows in it.

We assume that the stream, or at least a precursor of that stream, formed the valley in which it flows. This includes misfit streams (streams in valleys larger or smaller than are suggested by the current stream), although the size of the stream may be greatly altered. The riverine system is restricted to the lower portions of the valley. The riverine system may be as small as a meter or so across or as extensive as the Mississippi Valley and extends for hundreds or thousands of kilometers. Regardless of the size of the area, these systems are the result of a common set of fluvial processes. Riverine systems respond to, and have resulted from, hydrologic input from all parts of their upstream drainage basin and to a degree from their downstream basin. The hydrologic input and kinetic energy resulting from landscape relief creates the geomorphic features defining the riverine system; at any given point, the valley cross-section that results reflects the upstream and downstream conditions over time. Some features are quite transitory, such as those on the lowest floodplains with the youngest soils. Higher terraces are progressively more stable and have older soils (Daniels et al. 1971).

Floodplains are often the most obvious geomorphic feature of the riverine system and are a direct result of fluvial processes (Ritter 1979). The same processes also contribute to other associated landforms, including levees, oxbows, meander scars, bars, sloughs, and backswamps. (See Leopold et al. 1964 and Allen 1970 for more detail on these landforms.)

Fluvial Processes

Fluvial processes are driven by the kinetic energy of flowing water. Kinetic energy is derived from the elevation and gradient of the streams within their watershed. The watershed energy depends on overall relief from the highest portion of the watershed to the outlet. Within any given watershed, the fluvial processes depend on local relief factors and on the history of the stream itself; thus, the type and magnitude of the processes that occur in riverine systems vary depending on the location within the system. The basic fluvial activities in watersheds include runoff, landslides, channel erosion and deposition, and flood basin erosion and deposition. The material processes primarily consist of sediment entrainment, sediment transport, erosion, and deposition.

We generalize the fluvial processes within a watershed that form the valley and the landscape above the floodplain by using a hypothetical landform sequence based on the backwearing erosional model (Ruhe 1975). The first stage is a youthful stage without an integrated drainage network. This stage has high water tables and usually numerous wetlands, such as the prairie potholes or other closed depressions in a young glacial terrain (Figure 12.1a). The headward erosion of a transgressing channel has potential energy derived from the stream gradient and its water volume. Extra water volume derived from local wetlands and natural stream or channel drainage results in increased relief and extra water volume, which creates additional kinetic energy resulting in channel downcutting (Figures 12.1a through c) (Nash 1996). The result is a loss of wetlands and a lowered water table. Eventually, some base level of downcutting is reached. Base level is a point of severe resistance to downcutting due to a water table (in some cases sea level), indurated rock, or similar phenomena that resist denudation. The edges of the valley can slowly be altered by erosion of footslopes over long periods of time and by the backwearing of the valley edge (Figures 12.1d and e).

The factors responsible for the development of riverine systems (a stream forming landforms in its own alluvium) can be divided into two processes (channel process and floodplain process), each of which results in a specific floodplain type: an accretion floodplain from channel processes and an over-the-bank floodplain from flooding processes.

Valley widening and initial floodplain formation are a direct result of flowing water contained in the river channel (Langbein 1964; Allen 1970, p. 128). Water in the channel flows at varying velocities. When the water is moving slowly, deposition of the sediment occurs, often forming a point bar. On the opposite bank where the water is moving more rapidly, erosion occurs, resulting in the deepening of the channel and formation of pools. A line through the deepest sections of the channel is referred to as the thalweg (Figure 12.2). Typically, the thalweg does not remain in the center of the channel; instead, it migrates from side to side giving rise to a lateral component of the stream channel (Leopold et al. 1964). The shifting of the thalweg and subsequent differential erosion/deposition leads to the formation of meanders. The formation of meanders is related to the dynamic energy of the flowing water, the channel's slope, and its sediment load (Figure 12.3).

When a river meanders, its channel will stay within the confines of the valley. The resulting erosional and depositional processes result in cut and fill, deposition and erosion, or accretion that will cross the valley from edge to edge. Accretion can result in a scroll-shaped pattern of sediments (Figure 12.3) that form across a flat plain near a current stream (Leopold et al. 1964; Allen 1970; Hickin 1974). Generally with time, the accretion system migrates laterally across the valley several times, creating a seasonal floodplain (Leopold et al. 1964).

The accretion floodplain is entrenched in a larger system that is flooded less frequently but with larger flood events (floodplain processes rather than channel processes). Periodic inundation of the valley occurs during times when the volume of water in the stream exceeds the capacity of the channel (e.g., flooding during rainfall runoff and snow melt). This results in floods where erosional and depositional processes combine (over-the-bank flooding). These events deposit coarse sediments at the edge of the stream and create

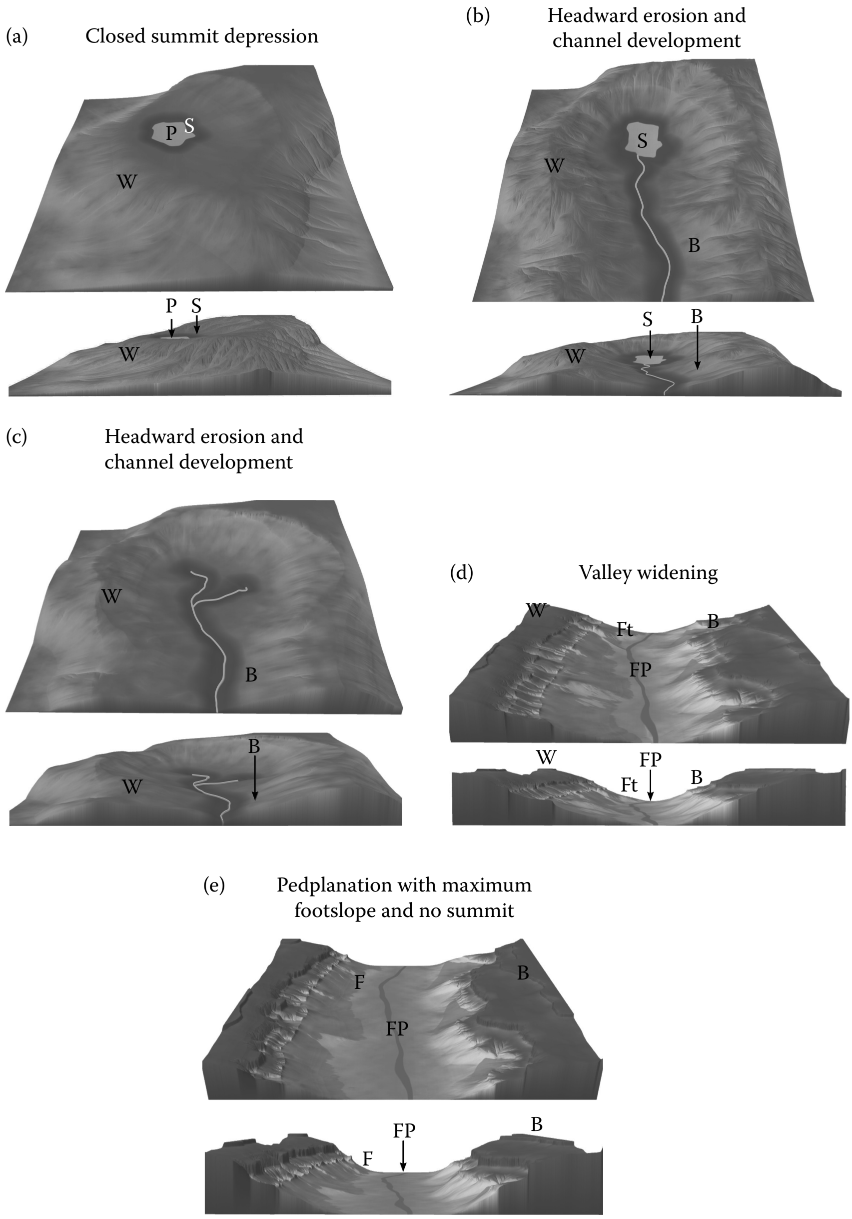


FIGURE 12.1
(a) Closed-summit depression illustrating a high water table. No channel or floodplain development has yet occurred. W is well drained; S is somewhat poorly drained; and P is poorly or very poorly drained. (b) With the invasion of either an artificial or natural channel, the water table is lowered and only somewhat poorly drained soils (S) in this case remain in the original wetland. An incipient or immature backslope is started (B). (c) The channel is well established and has down cut to a degree that the water table is only effective near the base of the backslope (B) with possible transient seeps above. The summit soils (W) are dominant and all are nearly well drained. (d) Backwearing or retreat of the backslope (B) into the upland summit (W) continues and the valley down cuts to a water table or any other restriction (base level). The valley widens (FP) based on the size of the stream. An erosional footslope or pediment starts (Ft). (e) Footslope (F) continues to develop at the base level created by the floodplain (FP). The backslope (B) retreats into the summit, and often, the summit is just coalesced on backslopes. The water table is high in the toeslope or floodplain and high in the lower footslope.

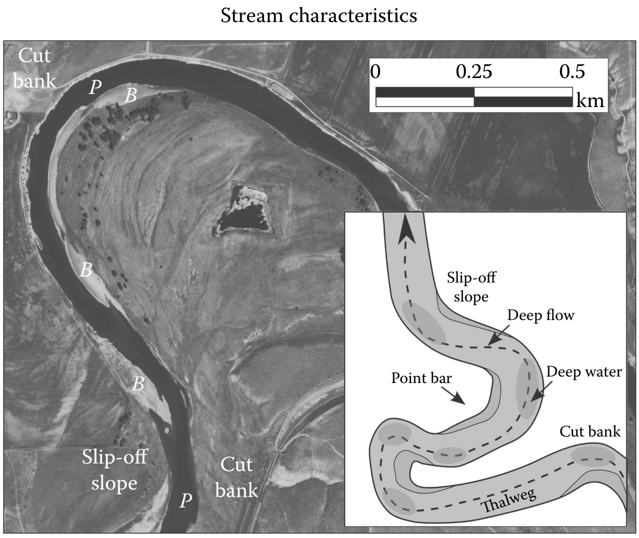


FIGURE 12.2
Essential parts of a stream or river channel. *P* = pool, *B* = point bar.

a local topographic high called a *natural levee.* Behind the natural levee, away from the stream, a landform called a *backswamp* develops (Figure 12.4). Backswamps are the largest and most extensive of riverine wetlands. The combination of these features is referred to as the over-the-bank floodplain (Figure 12.4). The water leaving the channel during flood stage often downcuts through the levee, creating a crevasse or a cut in the levee. The water then flows onto the floodplain from the levee, depositing coarser materials (splay deposits) (Figure 12.4). On the lower Mississippi, these distinct landforms are often named for the year they were first noted; cutoffs that produce oxbow lakes are similarly named.

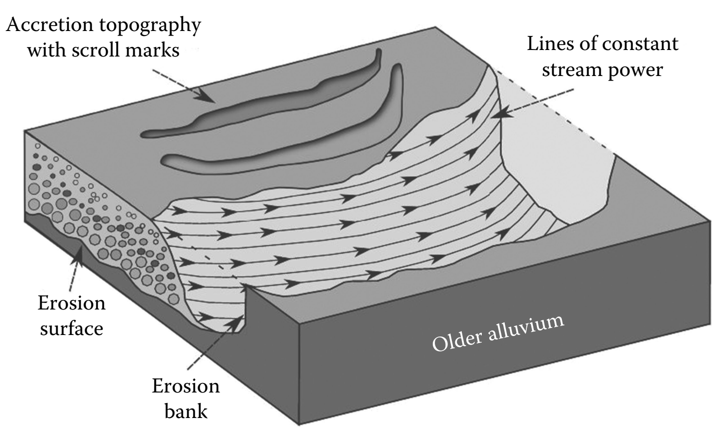


FIGURE 12.3
The thalweg creates strong water velocities on the outside that create a cutbank and fill on the inside (point bar). The point bar has coarser sediments deep, and fine upward, reflecting the decreasing energy with shallow water. Minor depressional wetlands with crescent shapes occur in accretion floodplains. (Adapted from Allen, J. R. L. 1970. *Physical Processes of Sedimentation.* American Elsevier Publishing Company, New York.)

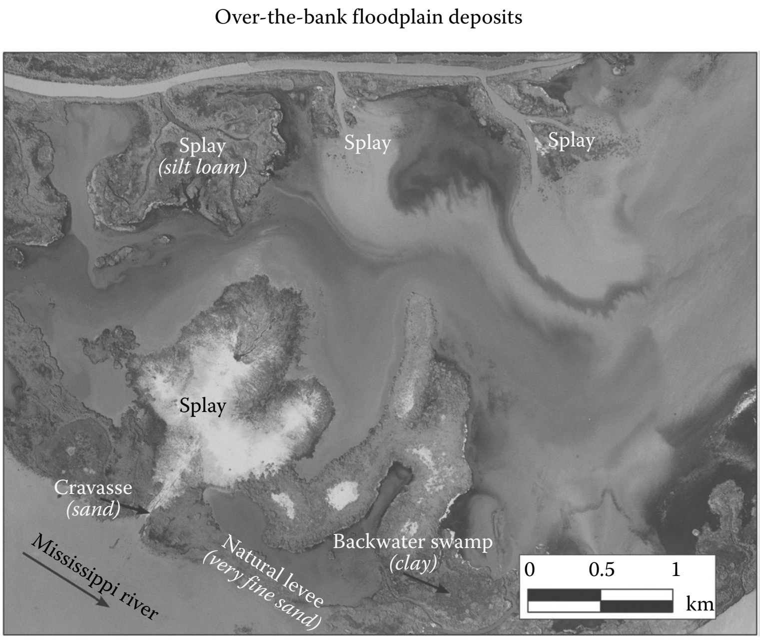


FIGURE 12.4
Map view of a small portion of a floodplain with over-the-bank landforms of various textures and elevations. The natural levee has the highest elevation, and backswamp is the lowest of these features.

Hydrologic Processes

Gaining and Losing Streams

In general, channels in riverine systems receive water that is discharging from the groundwater in the floodplain and are termed *gaining streams* (Figure 12.5a) (Todd 1980). Permanent streams are gaining streams. At high water, however, the channel yields water to the floodplain (recharge water) or releases flood water (surface water) (Figure 12.5b). In semiarid areas or in the upper reaches of streams in more humid areas, the stream may be a "losing" stream or a stream that gathers water above and recharges the water to groundwater (Figure 12.5c). Soils associated with losing streams are often unmodified sediments that classify as Fluvents. Fluvents are Entisols or recently formed soils that lack much horizon development. More soil development and features associated with hydric soils may be expected in soils associated with gaining streams.

Surface and Throughflow Water

Stream flow can be derived from two sources: surface runoff and base flow (Figure 12.6). Surface water flow is highly variable and travels rapidly to the stream. Surface runoff occurs during periods of high precipitation or snow melt. Water in excess of the soil's storage capacity results in overland flow to streams. In flood conditions, the stream channel's capacity to carry water is exceeded, and the stream channel overflows onto the floodplain. Wetland drainage and loss of natural vegetation often increase overland flow and

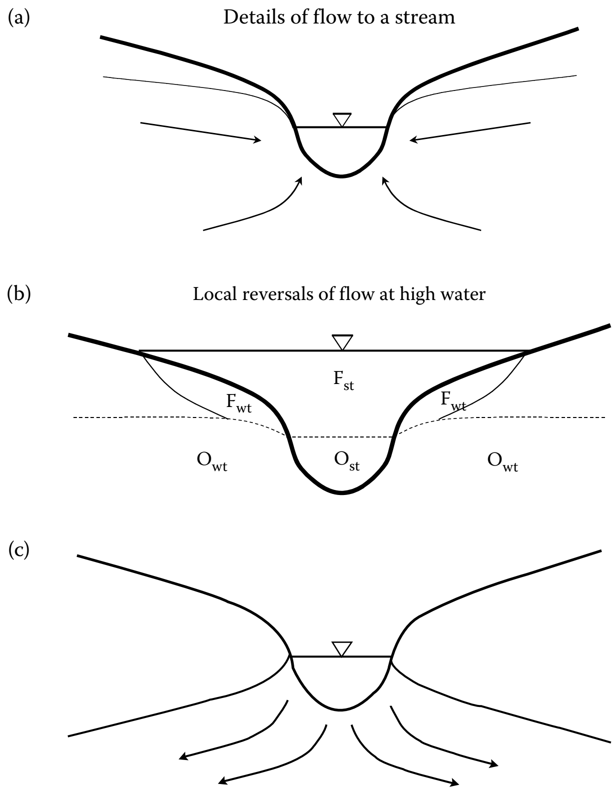


FIGURE 12.5
(a, top) Groundwater flow directions in a gaining stream. Note that the water level in the stream (inverted triangle) is relatively lower than the adjacent water table and that flow direction is toward the stream. (b, middle) Groundwater flow directions in a flooding stream. The water level in the flooded stream (F_{st}) is higher than the original stream level (O_{st}). During the early stages of flooding, the zone of saturated soil beneath the flood waters (F_{wt}) may not extend completely down to the original water table (O_{wt}). This may result in areas of entrapped air and unsaturated soil during a flood event. (c, bottom) Groundwater flow directions in a losing stream. Note that the water level in the stream (inverted triangle) is relatively higher than the adjacent water table.

flooding (Leopold et al. 1964). Excess precipitation or rapid snow melt in one part of a basin frequently results in flooding downstream. Water from flooding is stored in floodplain depressions, such as oxbows, and in backswamp landforms, creating wetlands. Evapotranspiration and slow groundwater release to the main stream occurs over time. An intermediate condition related to throughflow on slopes results in reflow or saturation of the soil and release or discharge of water at the base of slopes. The edge of floodplains often grades from a slope wetland to a riverine wetland.

Base Flow (Groundwater)

Base flow is that portion of the stream flow that is the result of groundwater discharge. Permanent streams are "gaining" streams that receive a steady influx of groundwater. Much of the groundwater is first discharged at floodplain soils, however, before being forwarded to the stream in its channel either by some groundwater movement or via yazoo streams on the floodplain. Many backswamp landforms develop streams that flow parallel

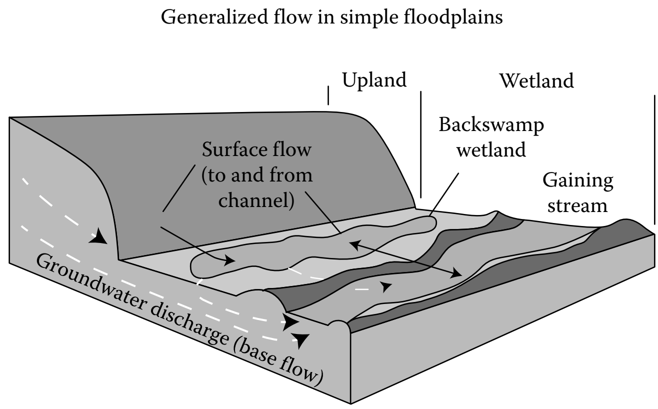


FIGURE 12.6
Surface and base flow into streams.

to the trunk stream for long distances. These are called *yazoo streams* after the Yazoo River in Mississippi, which flows in this manner before taking a sharp turn and discharging into the Mississippi River.

The general trend of both regional and local groundwater flow is from the uplands to the river valley (Gonthier 1996) (Figure 12.6). These waters discharge in the floodplain (backswamp-focused discharge) or more frequently at the valley edge (valley edge-focused discharge). These discharge areas are termed *seeps* (Figures 12.7 and 12.8). Many seeps are actually slope wetlands with organic soils, as illustrated in Figure 12.7 (Brinson 1993). Water discharging at the valley edge may be consistent enough for the formation of an organic soil (Histosol). The water at such discharge points is mineraltrophic (mineral rich) and has all the ingredients to form a fen or a graminoid-dominated, high-base wetland with organic soils. Many of these fens are slope wetlands or, as illustrated in Figure 12.7, a combination of slope and riverine wetlands; such wetlands often have slopes steeper than 4%. Malterer et al. (1986) describe a fen slope–riverine wetland combination, which had an organic thickness greater than 1 m and a riverine wetland with more than 4 m of muck.

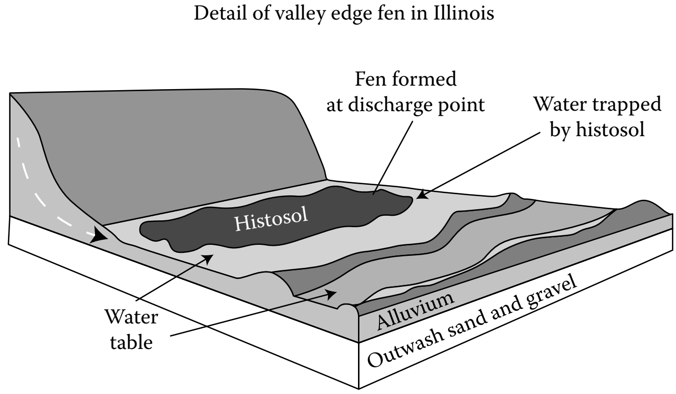


FIGURE 12.7
Discharge at the valley edge of the Des Plaines River in northern Illinois, creating a slope and riverine combination wetland.

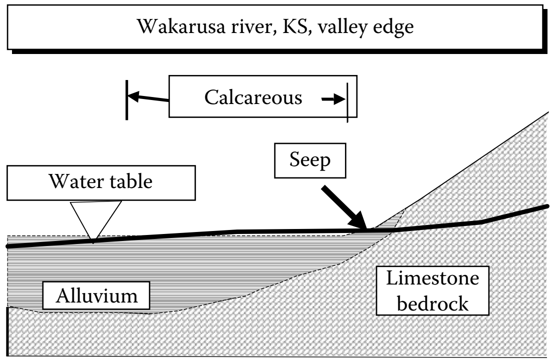


FIGURE 12.8
A detailed illustration of an edge seep with high amounts of Ca bicarbonate discharged into the soils. The limestone bedrock is slowly dissolved and added to the groundwater as Ca bicarbonate. At discharge into the soils, the temperature warms and the ability of the groundwater to hold CO_2 is reduced and calcite is precipitated.

Groundwater at discharge points often has a sudden alteration of its chemical regime. For instance, warming the water at the surface or evapotranspiration of water results in carbon dioxide (CO_2) being expelled from the solution. Loss of CO_2 removes the bicarbonate ions that are responsible for keeping soluble calcium (Ca) from precipitating as calcium carbonate ($CaCO_3$) (calcite). Some seeps have abundant calcite formed in this manner (Arndt and Richardson 1992; Almendinger and Leete 1998) as illustrated in Figure 12.8.

Deposits of ferric iron (Fe^{3+}) often form by oxidation of the groundwater as it is exposed to air at points of discharge. The mobile ferrous (Fe^{2+}) form of iron is reduced and moves with the groundwater. When exposed to the air, the Fe^{2+}-laden groundwater can often be observed as a plume of rusty-colored water seeping out of a bank (Figures 12.9 and 12.10) (Rhoton et al. 2002). Deposits of Fe large enough to mine, called bog iron, were common along streams near the Atlantic Coast. Many were exploited as an iron ore during the colonial period. Sediment and soil beneath these oxidized surfaces remain reduced.

FIGURE 12.9
Example of oxidized iron entering a stream through a seep at the valley edge. (Photo from Lafayette County, MS.)

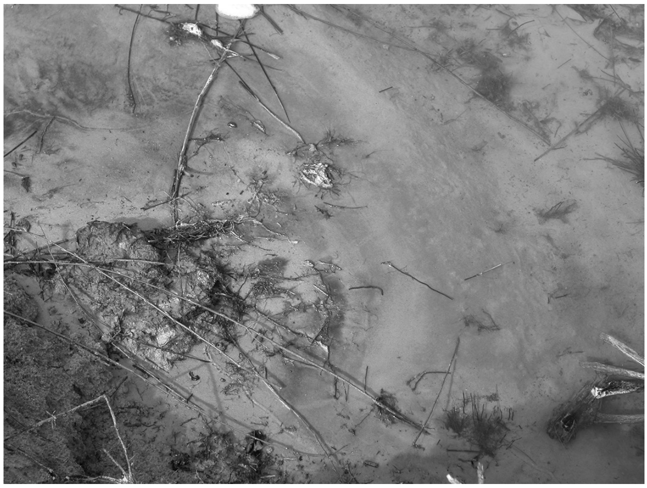


FIGURE 12.10
(**See color insert.**) Example of oxidized iron entering a stream through a seep at valley edge. (Photo from Chowan Co., NC, provided by David Lindbo.)

Geomorphic Features of Floodplains

A presently forming floodplain will often have accretion topography, or exhibits entrenching, which results in high slopes above the outside of the river's meanders and low "slip-off slopes" on the inside of meanders (Figure 12.3). In nonentrenched streams, an over-the-bank floodplain has usually formed as a result of periodic inundation (Ritter 1979). The size and shape of the floodplain is the result of variable stream flow and sediment load throughout the drainage basin, which has responded to factors unique to the hydrology such as climate, time, topography, and geology. The system's sediments are composed of alluvium transported and deposited by the stream or river. These sediments may have been transported from the system's headwaters or derived locally from channel erosion and subsequent deposition. At the edge of the floodplain, colluvial sediment may be present, which is derived from the uplands at the river valley edge. The modern floodplain mitigates the effects of the flood by acting as storage for both water and sediment. As such, the floodplain has both form and function within the riverine system (Wolman and Leopold 1957).

Although floodplains appear to be topographically simple, they are composed of a variety of features (Figure 12.11), some of which can be anthropogenic (e.g., mill ponds and their deposits). At the valley edge where the uplands and valley floor meet, it is common to observe colluvial material that has been deposited by mass movement off the valley sides. Colluvial material will grade into alluvial material toward the valley center. Numerous deposition zones are common within these alluvial sediments and include: (1) coarse-textured lag deposits in which the finer sediment has been selectively removed, (2) poorly sorted fill resulting from bank and channel collapse, (3) coarse-to-medium-textured near-bank and point bar sediments deposited as stream energy decreases, and (4) finer-textured overwash sediments deposited as flood waters flow out over the floodplain. The sequence of textures in Figure 12.12 is based on soils from the Meherrin River floodplain in Virginia (Richardson and Edmonds 1987). The textures represent an energy of

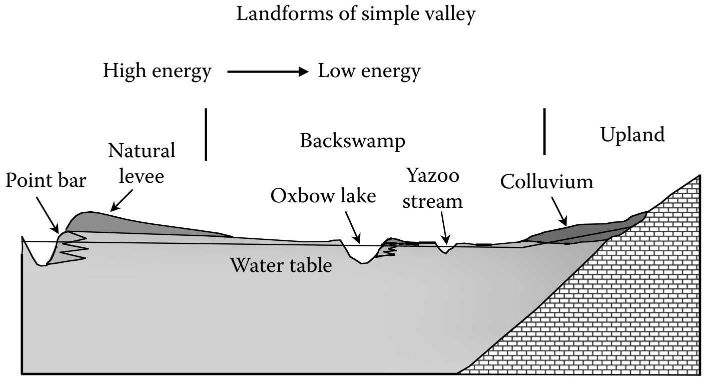


FIGURE 12.11
Cross-section of floodplain features indicating relative energy of deposition. An older, no longer accreting point, bar and the natural levee occur adjacent to the oxbow lake. Within the floodplain, alluvium depositional facies are present but are not shown for the sake of clarity.

deposition geosequence, from the coarsest (left by the strongest currents) to the finest (e.g., backswamp clays) materials deposited in the most quiet conditions.

The overall topography of the floodplain becomes less flat and smooth in the area adjacent to the active river channel because of the active fluvial process of channel erosion (an example of *accretion topography*: point bar area in Figure 12.12). Within the channel, point bars are likely to be seen in areas where flow is slower, and are likely to be coarser textured than the surrounding features (Figure 12.12). As the river meanders over time, point bars will be reworked into low ridges and troughs referred to as *meander scrolls.* The low ridges are rapidly vegetated and, in some instances may act as a channel bank. It is common for a series of these scrolls to occur across the floodplain. The trough between the ridges, sometimes referred to as a slough or chute, may eventually fill with fine-grained material, but it is usually distinctly wetter than the surrounding meander scrolls. Such

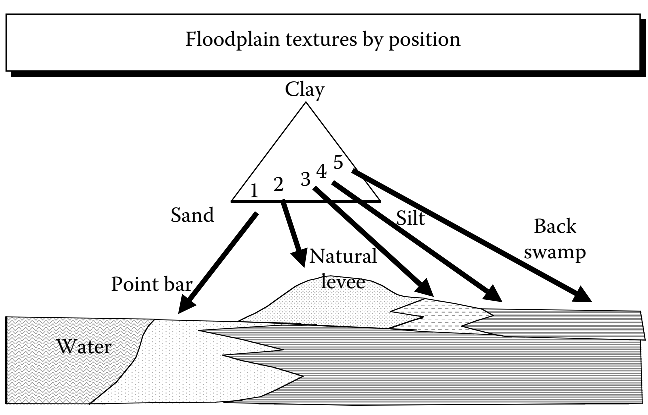


FIGURE 12.12
Texture of soils along the Meherrin River floodplain in Greenville County, Virginia, that form a geosequence based on the decreasing energy of deposition. (Data from Richardson, J. L. and W. J. Edmonds. 1987. *Soil Sci.* 144: 203–208.)

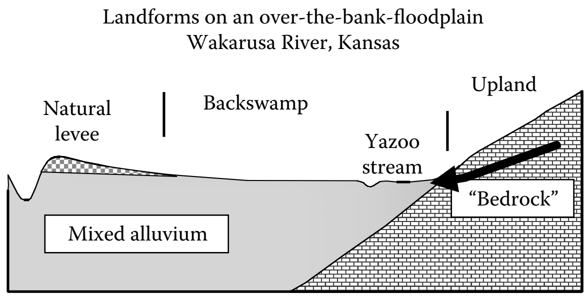


FIGURE 12.13
Cross-section of a valley with a distinct over-the-bank floodplain and an entrenched stream. The Wakarusa River near Lawrence, Kansas, has a distinct natural levee well above the river channel, which grades into the backswamp landform with a yazoo stream.

a sequence may be stable for a number of years and may have a scroll-like appearance when viewed from above, hence the term *meander scrolls.* An example is shown in Figures 12.13 and 12.14. As noted by Allen (1970), the point bar deposits are coarser at depth and fine upward.

Natural levees are, in some regards, similar to meander scrolls, because they appear as low ridges adjacent to the channel. Unlike the meander scrolls, however, natural levees are formed during flood events as the flood water flow decreases after the water leaves the channel and coarser-suspended sediment is deposited. These ridges grade into the low flat areas of the floodplain referred to as the backswamp (Figure 12.13). The Wakarusa River near Lawrence, Kansas, has a distinct natural levee well above the river channel. This river is deeply entrenched into its channel such that low-flow water levels are well below the floodplain. Despite this deep entrenchment, the natural levee gradually grades into the floodplain and the backswamp landform. The backswamp is composed of fine-grained sediment. Periodically, a levee is breached during large floods. A channel is cut through the levee and forms a splay deposit, which consists of sediments that are finer than the levee but coarser than those of the backswamp (Figure 12.4). Most of the backswamp qualifies as having hydric soils.

As the river channel migrates laterally through the floodplain, sections of the channel may be cut off from the main river (Figure 12.14). These cutoffs form oxbows or oxbow lakes if they remain filled with water. Over time, the oxbows will fill with fine-grained sediments, and perhaps organic matter (OM), particularly if the oxbow remains water filled with little or no turnover of the water. OM (leaf litter, woody vegetation, etc.) will decompose slowly in the anaerobic water. These features, even after filled in, remain visible for many years.

Soil Distribution

The complex nature of the riverine system, with backswamps, meander scar sloughs, and oxbows, results in a patchwork of environments based on the topographic position, distance from the channel, relation to the flood stage, and texture (Daniels and Hammer 1992;

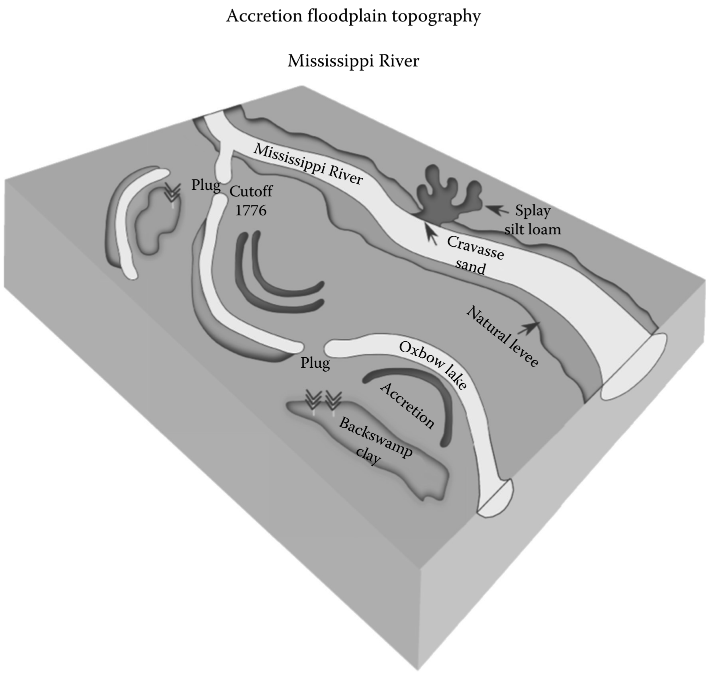


FIGURE 12.14
Map of an accretion topography with an oxbow wetland and lake created during 1776 on the Mississippi River.

Daniels et al. 1999). The variety of sedimentary deposits gives rise to a complex distribution of soils across the floodplain (Figure 12.15). The nature of both lateral and horizontal changes in sediment deposition and texture result in soil series with a large degree of variability (Leab 1990). This combination makes standard soil survey maps of floodplains difficult to interpret, because they cannot adequately delineate all the details of the area at the scale of mapping used. Therefore, detailed investigations are often required to fully characterize these soils; detailed investigations are certainly needed for hydric soil delineation. Anthropogenic landscape disturbance can greatly alter sediment distribution rates and patterns in such systems (Ricker et al. 2012, 2013).

Wetlands and Hydric Soils

A wetland is defined as having wetland hydrology, a predominance of hydrophytic vegetation, and the presence of hydric soils. A hydric soil is identified based on its morphology or on the extent of flooding or ponding that the soil is subject to over a given time span (USDA-NRCS 2010). Some soils may not meet the current hydric indicators (USDA-NRCS 2010); yet, be hydric by virtue of receiving enough water to be ponded or flooded; in such cases, new hydric soil indicators should be identified.

Areas that are periodically inundated with floodwaters for a significant period during the growing season may be considered jurisdictional wetlands if there is a presence of hydric soil, hydrophytic vegetation, and evidence of wetland hydrology (Environmental

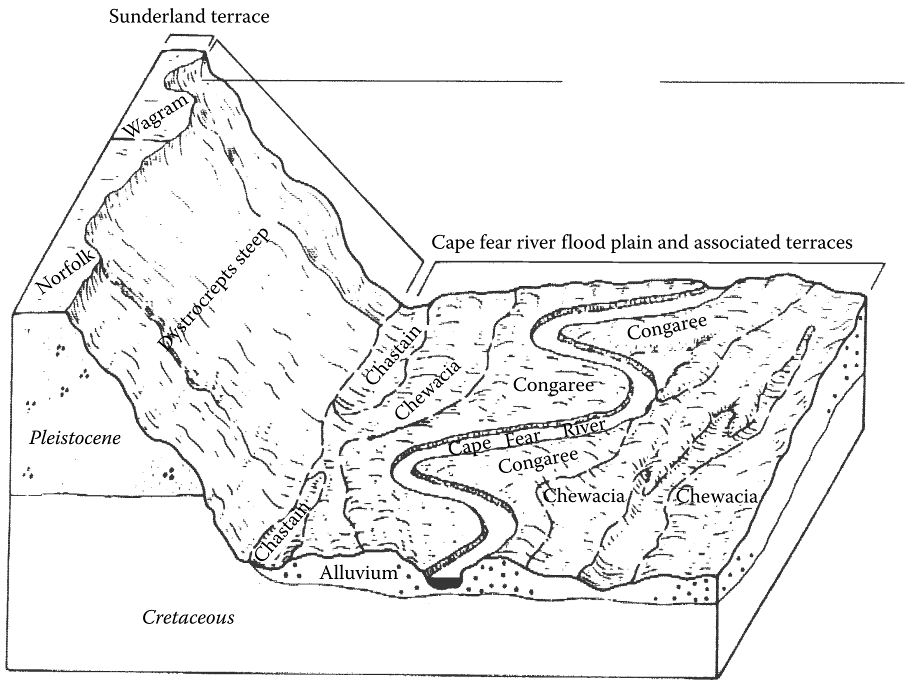


FIGURE 12.15
Soil distribution within the Cape Fear floodplain and associated terraces. (From Leab, R. J. 1990. *Soil Survey Report of Bladen Company, North Carolina.* USDA, SCS, U.S. Government Printing Office, Washington, DC.)

Laboratory 1987). A hydric soil is "a soil that formed under conditions of saturation, ponding, or flooding long enough during the growing season to develop anaerobic conditions in the upper part" (Federal Register 1994). The required time that saturation and anaerobic conditions must exist for a hydric soil is at least 14 continuous days (NTCHS 2007). In a broad sense, this potential hydrologic footprint and time frame covers much of the active floodplain, within which can be found a range of geomorphic and hydrologic zones. These differences result in a gradation of soil morphologies and hydric soil locations across the floodplain that can be divided into two groups: those related to the channel and/or flood events, and those associated with seeps at the valley edge (Figure 12.16).

Wetland and hydric soil distribution across the floodplain are directly related to geomorphic features; hydrophytic vegetation is common where inundation is frequent (Veneman and Tiner 1990). In general, the higher areas, such as natural levees and larger scroll ridges, contain soils that are nonhydric, whereas the sloughs and backswamp areas, and features within the channel contain soils that are hydric. Farther away and higher in elevation than the active floodplain are the older terraces. Like the active floodplain, only the soils in low-lying areas are likely to be hydric. Oxbows are likely to contain hydric soils as well. Any of the natural levees or other ridges remaining near the oxbow are likely to be nonhydric. In addition to the fluvial and channel-related hydric soils are those associated with groundwater seeps at the valley edge. These occur when the underlying stratigraphy allows for lateral groundwater flow that eventually surfaces at the valley edge (Figures 12.6 through 12.8).

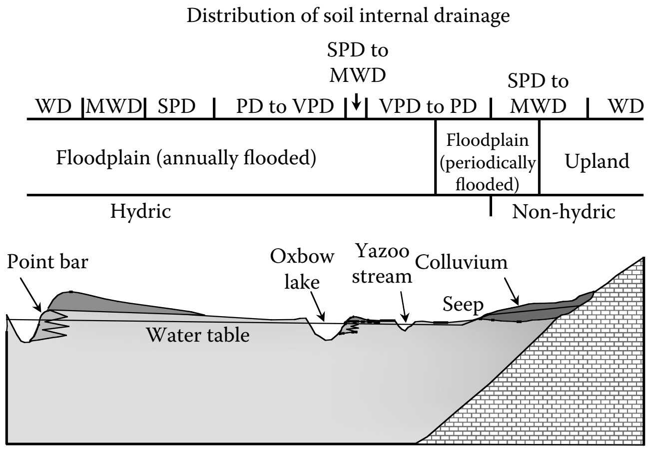


FIGURE 12.16
Cross-section showing hydric soil relations on a floodplain and at the valley edge. Within the active (annual) floodplain, well-drained (WD), moderately well-drained (MWD), and somewhat poorly drained (SPD) soils occur; yet, due to annual inundation, these are considered hydric soils along with the poorly drained (PD) and very poorly drained (VPD) soils. At the valley edge, a seep in colluvium, a PD, or VPD soil may occur and qualify as a hydric soil.

Hydric Soil Indicators

The distribution, delineation, and formation of hydric soils are of particular interest. Several field indicators of hydric soils directly deal with hydric soils on floodplains (USDA-NRCS 2010, Chapter 8). It is probable that nearly all indicators occur in hydric soils of riverine systems; however, a few of them are common to many riverine wetland types. These include A2: histic epipedon; A4: hydrogen sulfide; A5: stratified layers; A11: depleted below a dark surface; S5: sandy redox; S7: dark surface; S9: thin dark surface; F3: depleted matrix; F12: iron/manganese masses; and F13: umbric surface. Two regionally extensive indicators are F17: Delta Ochric (Mississippi River Delta) and F19: Piedmont Flood Plain Soils (U.S. Mid-Atlantic and Southern Piedmont). Indicator TF2: Red Parent Material is an indicator common to Piedmont, and is often difficult to document in the field due to the inherently red soil matrix.

Young riverine wetlands, while perhaps saturated for extensive periods, may not yet exhibit a strong morphology typical of extensive reduction and OM accumulation. As hydric soil indicators have been developed, such conditions have been taken into account as in the case of F12, which stipulates that only “40% or more of the matrix” needs to have a chroma of 2 or less. Essentially, this recognizes that riverine wetlands in active floodplains frequently do not get the amount of iron depletion (reduction) that other wetland landscapes are subjected to, or at least it is not visible. This can be because the young soils contain little-to-no leaf litter accumulation, have low carbon contents, or receive extra iron via groundwater discharge that is precipitated as bog iron. The young soils may be formed from oxidized sediments and are deposited quickly before reduction occurs. In indicator F12, the thickness requirement for the zone or horizon is waived if the indicator is found in the mineral surface layer. This recognizes that addition of the sediment during a flood may result in the shallow or solely surface development of the indicator, because insufficient time has passed for deeper pedogenesis. Additionally, because organic carbon (OC) is critical to soil reduction and the formation of low chroma colors, both reduction and low chroma colors will be absent if sufficient OC is not present. The extra iron in groundwater that discharges may reconstitute that high chroma coloration in these soils.

Difficulties exist in identifying hydric soils on floodplains and riverine systems. This can be due to the duration of relationship between oxygenated floodwaters and groundwater, reduction versus oxidation, lithochromic colors inherited from the parent material, low carbon content, and relict morphologies, especially in soils located on terraces. In fact, terrace soils may have formed in saturated conditions possessing redoximorphic features, but may currently exist in well-drained conditions (relict) because of entrenching. The actual creation of the terrace by entrenching streams leaves a portion of the terrace better drained than it was originally; these are the areas with relict wetness conditions.

Flood Relationships and Redox Status

Rainwater is generally oxygenated, and rivers swollen by rainwater are usually oxygenated as well. During a flood, the water being added to the floodplain must become depleted of oxygen before the soil becomes reduced and begins to form redoximorphic features and related hydric soil indicators. For oxygen to be depleted, the water must stagnate. This occurs more often in low-energy environments where the water is not flowing (Veneman and Tiner 1990; Faulkner et al. 1991). This relationship has been observed in a study of bottomland hardwood forests in the lower Mississippi River Valley. Soils in areas with higher hydrologic energy (active floodplains) developed anaerobiosis for shorter periods than soils in quieter, backwater areas (ponded conditions). Similar hydric soil morphologies (redoximorphic features) were observed in soils that were reduced and saturated for nearly 100% of the growing season and in those that were reduced and saturated for as little as 10% of the growing season. It seems plausible that hydric soil morphologies reflect a critical duration of anaerobiosis and change slowly after that period has been attained (Faulkner et al. 1991).

Another aspect of landscape position influencing the hydric status of the soil is the type of microtopography (Veneman and Tiner 1990). Closed-drainage areas (depressions without surface flow outlets) on the upper floodplain, or terraces, tend to have hydric soils, while areas with open drainage, even on the active floodplain, do not. Thus, if aerated floodwaters are exchanged in the active floodplain, the soil water could remain aerobic even when flooded. Hydrophytic vegetation was observed throughout the active (young) floodplain despite some areas having soils that lacked hydric soil morphology.

Air may become entrapped as floodwaters saturate soil from the surface down. Such an occurrence has been observed on the floodplain of the Connecticut River (Chase-Dunn 1991). Some of the soils are assumed to remain aerobic as suggested by escaping air bubbles and the rapid fall of the water table after inundation. Further indication of aerobic conditions comes from *in situ* measurements indicating that reducing conditions do not always occur after flooding. Aerated floodwater flowing down macropores (such as root or worm channels) is the most probable explanation for the aerobic conditions observed in the soil. These macropores allow for the rapid exchange of aerated water or air into the soils. Measurement of conditions in a macropore would not reflect the true conditions in the soil matrix (Mukhtar et al. 1996). Additional research is needed in this area.

Lithochromic and Relict Colors

Two hydric soil indicators for floodplains (TF2: Red Parent Material, and F17: Delta Ochric) were identified specifically because of the problem of soil colors inherited from the parent material, also referred to as *lithochromic colors.* In instances where the sediment accumulating on the floodplain comes from an area where the bedrock and/or soils are red (7.5YR or redder), then, it too will have a red coloration. This color, likely due to hematite coatings, is more resistant to changes due to redox status, and persists longer in the soil. The result is a high chroma soil in a reducing environment. Such a situation was observed in a transect across the Red River floodplain in Louisiana (Faulkner et al. 1991). The alluvium in this floodplain was derived from Permian red bed parent material and was observed to be resistant to color change. For this reason, it was not possible to compare soil morphologies to redox conditions in the soils.

Alluvium derived from gray or low chroma parent materials presents a contrasting dilemma. Such is the case in the Connecticut River Valley where soils with similar morphologies have different degrees of saturation and reduction (Veneman and Tiner 1990; Chase-Dunn 1991). In this situation, two profiles of the Limerick series (coarse–silty, mixed, nonacid, mesic, and Aeric Fluvaquent), both having similar morphologies, were monitored for saturation and reduction. The results showed one to have hydric soil conditions, while the other did not. Furthermore, both soils had similar OC and free iron contents. Both these soils were young, frequently flooded soils (possibly <20 years old in the upper 45 cm), and lacked the redoximorphic features that would help identify them as hydric soils. There was a similarity in color between the C horizons (or mineral strata) and the upper profile (0–30 cm). The lithochromic influence with matrix chromas less than 2 may suggest that a soil is saturated and reduced when it is not. As a result, these studies suggest that monitoring may be necessary to confirm wetland and hydric soil status.

One solution for identifying potential-problem hydric soils is to use a color change propensity index (CCPI) (Rabenhorst and Parikh 2000). This CCPI index helps one to discern color changes in soils under reducing conditions via the use of a digital colorimeter and soils that have been treated with dithionite–citrate–bicarbonate for a period of time at a certain temperature. On the basis of a limited data set of the United States, Rabenhorst and Parikh (2000) have proposed that nonproblematic soils have a CCPI above 40 while problematic soils have a value at or below 30.

OM and Temperature Relationships

Reducing conditions are crucial for the formation of redoximorphic features common in identifying hydric soils (see Chapters 4 and 7). For reduction to occur, sufficient OM must be present in the soil. One study of a constructed floodplain suggests that redoximorphic features are formed during short periods of inundation (after one event) when soil OM is >30 g kg^{-1}, and are not found in soils with an OM concentration of <15 g kg^{-1} (Vepraskas et al. 1995). The first features formed were small and difficult to see with the naked eye, but with time, their abundance and size increased.

A study of some alluvial soils in the Puget Lowlands, WA, indicated that aerobic conditions persisted whether the soil was saturated or not (Cogger and Kennedy 1992; Cogger et al. 1992). It was further concluded that, despite the overall aerobic conditions, some microsite-reducing conditions did occur. The study concluded that approximately 10% of the observed field variation was due to the inherent variability of the electrodes, while the remainder was attributed to microsites in the soil. This conclusion was based on several field and controlled laboratory investigations. The lack of the overall reducing conditions was attributed to two factors: first, low levels of available carbon present; second, low temperatures during the period of saturation (6°C on the surface). It was demonstrated that reducing conditions occurred only after 3–6 months of saturation in soils with low OM and at low temperatures. The combination of low OM and low temperatures inhibits or delays reducing conditions and redoximorphic feature formation.

Relict Features

Changes in the overall drainage due to variation in stream drainage (either natural or altered by humans) may result in the current morphology and/or chemical composition reflecting historic rather than current conditions. This can be observed on upper terraces associated with active floodplains or in active floodplains of watersheds experiencing extensive land use change (Noe and Hupp 2005; Walter and Merritts 2008; Ricker et al. 2013). Some of these soils will retain their hydric morphology if insufficient time has passed to allow for pedogenic processes to reflect current conditions. Retention of hydric morphology can be influenced by OM and temperature, as indicated above, as well as by available Fe/Mn (if the soil was depleted in Fe, it is not likely to redden rapidly once drained).

Restoration of Riverine Wetlands

Riverine wetlands are perhaps the most important type of wetland to humanity due to dependence on them for supporting aspects of navigation, food, power production, development, flood control, and recreation (Smith et al. 2008). Today, their protection is essential for flood control and wildlife habitat, and these goals have resulted in numerous efforts at restoring the degraded function. Assessment of ecological integrity is typically the first step in restoration (Jungwirth et al. 2002; Jacobs et al. 2010) and can provide valuable knowledge of how different a degraded wetland system is. The key to riverine wetland restoration is the return of the river to "natural" flow patterns whereby flood pulsing is common with overbank discharge returning a variety of particle sizes, OC, and nutrients. Additionally, the redevelopment of a river's major spatial elements is desired (Ward et al. 2002). This cycle of pulsing will result in a change in river morphology that in time results in an equilibrium reflective of current up- and downstream watershed processes (Smith et al. 2008), but does not necessarily increase the prevalence of riverine wetlands (Kroes and Brinson 2004). Disturbed riverine wetlands can have lower OC and higher silt and clay (Drohan and Brooks 2013); however, flow dynamics in recovering riverine systems are complex and can result in unpredictable patterns (Anderson and Mitsch 2006). The addition of OC during restoration can help return wetland soils to levels more common of lesser-disturbed areas (Stauffer and Brooks 1997; Bruland and Richardson 2004). In addition, OC additions may improve the recovery of organisms dependent on the OC and result in faster OC accumulation as a whole as plant production increases due to increases in nutrient and water-holding capacity. Theoretically, the OC effect could be influenced by the release of seed-bank vegetation with disturbance, an encroachment of invasive species during restoration, and the time of year restoration occurs. Finally, there has been much interest in the use or geoengineering of riparian zones and wetlands to help reduce agriculturally derived nitrogen (Mitsch et al. 2005, 2008), especially in river waters draining to the Gulf of Mexico. Engineered riparian and wetland systems in the Mississippi River Basin could result in 22,000 km^2 of created and restored wetlands, many areas potentially including riverine wetlands. Suggestions for the use of engineered wetland systems around the world to improve water quality have received much attention. Given the less likely alternative of changing farming practices to reduce nitrogen fertilizer (Donner and

Kucharik 2008), it would seem that the future of riverine wetland management could encompass a scale never before seen in history.

Summary

Hydric soils and wetlands are influenced by the same processes that combine to form the features common in riverine systems. In general, all these processes are related to water as it moves through the system either as surface or groundwater flow. The dynamic features of the floodplain landscape contain several challenges to hydric soil identification. The most notable of them are the influences of aerated flood waters, low OC, lithochromic colors, young sediments, and relict features. Many of these problems can be overcome through a detailed site evaluation and full understanding of the hydrology of the riverine system.

References

Allen, J. R. L. 1970. *Physical Processes of Sedimentation.* American Elsevier Publishing Company, New York.

Almendinger, J. E. and J. H. Leete. 1998. Regional and local hydrology of calcareous fens in the Minnesota River Basin, USA. *Wetlands* 18: 184–202.

Anderson, C. J. and W. J. Mitsch. 2006. Sediment, carbon, and nutrient accumulation at two 10-year old created riverine marshes. *Wetlands* 26(3): 779–792.

Arndt, J. L. and J. L. Richardson. 1992. Carbonate and gypsum chemistry in saturated, neutral pH soil environments. In R. D. Robarts, and M. L. Bothwell (Eds.) *Aquatic Ecosystems in Semi-Arid Regions*. National Hydrology Research Symposium Series 7. Environment Canada, Ottawa, Ontario, pp. 179–187.

Brinson, M. M. 1993. *A Hydrogeomorphic Classification for Wetlands.* Technical Report TR-WRP-DE-4. U.S. Army Corps of Engineers, Waterways Experiment Station, Vicksburg, MS.

Bruland, G. L. and C. J. Richardson. 2004. Hydrologic gradients and topsoil additions affect soil properties of Virginia created wetlands. *Soil Sci. Soc. Am. J.* 68: 2069–2077.

Chase-Dunn, C. K. 1991. *Soil Mottling as an Indicator of Seasonal High Water Table in Massachusetts Floodplain Soils.* MS thesis, University of Massachusetts, Amherst, MA, 125p. (Unpublished).

Cogger, C. G. and P. E. Kennedy. 1992. Seasonally saturated soils in the Puget Lowland. I. Saturation, reduction and color patterns. *Soil Sci.* 153: 421–433.

Cogger, C. G., P. E. Kennedy, and D. Carlson. 1992. Seasonally saturated soils in the Puget Lowland. II. Measuring and interpreting redox potentials. *Soil Sci.* 154: 50–58.

Daniels, R. B., S. W. Buol, H. J. Kleiss, and C. A. Ditzler. 1999. *Soil Systems of North Carolina.* North Carolina State University, Raleigh, NC.

Daniels, R. B., E. E. Gamble, and J. G. Cady. 1971. The relation between geomorphology and soil morphology and genesis. *Adv. Agr.* 23: 51–88.

Daniels, R. B. and R. D. Hammer. 1992. *Soil Geomorphology.* John Wiley & Sons, New York.

Donner, S. D. and C. J. Kucharik. 2008. Corn-based ethanol production compromises goal of reducing nitrogen export by the Mississippi River. *Proc. Natl. Acad. Sci.* 105(11): 4513–4518.

Drohan, P. and Brooks, R. P. 2013. Hydric soils across Pennsylvania reference, disturbed, and mitigated wetlands. In Brooks, R. P. and D. H. Wardrop (Eds.) *Mid-Atlantic Freshwater Wetlands: Advances in Wetlands Science, Management, Policy, and Practice*. Springer, New York, pp. 129–157.

Environmental Laboratory. 1987. *Corps of Engineers Wetlands Delineation Manual*. Technical Report Y-87-1. U.S. Army Corps of Engineers Waterways Experiment Station, Vicksburg, MS. http://el.erdc.usace.army.mil/wetlands/

Faulkner, S. P., W. H. Patrick Jr., R. P. Gambrell, W. B. Parker, and B. J. Good. 1991. *Characterization of Soil Processes on Bottomland Hardwood Wetland–Nonwetland Transition Zones in the Lower Mississippi River Valley*. Contract Report WRP-91-1. U.S. Army Corps of Engineers Waterways Experiment Station, Vicksburg, MS.

Federal Register. 1994. *Changes in Hydric Soils of the United States*. National Archives and Records Administration and U.S. Government Printing Office, Washington, DC, available at; http://www.gpo.gov/fdsys/pkg/FR-1994-07-13/html/94-16835.html (Verified, 3/10/2014).

Gonthier, G. J. 1996. Groundwater flow conditions within a bottomland hardwood wetland, Eastern Arkansas. *Wetlands* 16: 334–346.

Hickin, E. J. 1974. The development of meanders in natural river channels. *Am. J. Sci.* 274: 412–442.

Jacobs, A. D., M. E. Kentula, and A. T. Herlihy. 2010. Developing an index of wetland condition from ecological data: An example using HGM functional variables from the Nanticoke watershed, USA. *Ecol. Indicators* 10: 703–712.

Jungwirth, M., S. Muhar, and S. Schmut. 2002. Re-establishing and assessing ecological integrity in riverine landscapes. *Freshwater Biol.* 47: 867–887.

Kroes, D. E. and M. M. Brinson. 2004. Occurrence of riverine wetlands on floodplains along a climatic gradient. *Wetlands* 24: 167–177.

Langbein, W. B. 1964. Geometry of river channels. *J. Hydraul. Div. ASAE* 90: 301–312.

Leab, R. J. 1990. *Soil Survey Report of Bladen Company, North Carolina*. USDA, SCS, U.S. Government Printing Office, Washington, DC.

Leopold, L. B., M. G. Wolman, and J. P. Miller. 1964. *Fluvial Processes in Geomorphology*. W. H. Freeman, San Francisco.

Malterer, T. J., J. L. Richardson, and A. L. Duxbury. 1986. Peatland soils associated with the Souris River, McHenry County, North Dakota. *Proc. Natl. Acad. Sci.* 40: 103.

Mitsch, W. J., J. W. Day, L. Zhang, and R. P. Lane. 2005. Nitrate–nitrogen retention in wetlands in the Mississippi River Basin. *Ecol. Eng.* 24: 267–278.

Mitsch, W. J., L. Zhang, D. F. Fink, M. E. Hernandez, A. E. Altor, C. L. Tuttle, and A. M. Nahlik. 2008. Ecological engineering of floodplains. *Ecohydrol. Hydrobiol.* 8: 139–147.

Mukhtar, S., J. L. Baker, and R. S. Kanwar. 1996. Effect of short-term flooding and drainage on soil oxygenation. *Trans. Am. Soc. Agric. Eng.* 39: 915–920.

Nash, D. J. 1996. Groundwater sapping and valley development in the Hackness Hills, North Yorkshire, England. *Earth Surf. Process. Landf.* 21: 781–795.

National Technical Committee for Hydric Soils. 2007. The Hydric Soil Technical Standard. Hydric Soils Technical Note 11, http://soils.usda.gov/use/hydric/ ntchs/tech_notes/index.html

Noe, G. B. and C. R. Hupp. 2005. Carbon, nitrogen, and phosphorus accumulation in floodplains of Atlantic Coastal Plain rivers, USA. *Ecol. App.* 15: 1178–1190.

Playfair, J. 1802. *Illustrations of the Huttonion Theory of the Earth*. William Creech, Edinburgh.

Rabenhorst, M. C. and S. Parikh. 2000. Propensity of soils to develop redoximorphic color changes. *Soil Sci. Soc. Am. J.* 64: 1904–1910.

Rhoton, F. E., J. M. Bigham, and D. L. Lindbo. 2002. Properties of iron-oxides in streams draining the loess uplands of Mississippi. *Agronomy Abstracts* 17(4): 409–419.

Richardson, J. L. and W. J. Edmonds. 1987. Linear regression estimation of Jenny's relative effectiveness of state factors equation. *Soil Sci.* 144: 203–208.

Ricker, M. C., S. W. Donohue, M. H. Stolt, and M. S. Zavada. 2012. Development and application of multi-proxy indices of land use change for riparian soils in southern New England, USA. *Ecol. Appl.* 22: 487–501.

Ricker, M. C., M. H. Stolt, S. W. Donohue, G. A. Blazejewski, and M. S. Zavada. 2013. Soil organic carbon pools in riparian landscapes of Southern New England. *Soil Sci. Soc. Am. J.* 77: 1070–1079.

Ritter, D. F. 1979. *Process Geomorphology*. W. C. Brown, Dubuque, IA.

Ruhe, R. V. 1975. *Geomorphology*. Houghton Mifflin Company, Boston.

Smith, L. M., N. H. Euliss Jr., D. A. Wilcox, and M. M. Brinson. 2008. *Application of a Geomorphic and Temporal Perspective to Wetland Management*. Environmental Science and Biology Faculty Publications. Paper 82. http://digitalcommons.brockport.edu/env_facpub/82

Stauffer, A. L. and R. P. Brooks. 1997. Plant and soil responses to salvaged marsh surface and organic matter amendments at a created wetland in central Pennsylvania. *Wetlands* 17: 90–105.

Todd, D. K. 1980. *Groundwater Hydrology*. John Wiley & Sons, New York.

United States Department of Agriculture, Natural Resources Conservation Service (USDA-NRCS). 2010. Field Indicators of Hydric Soils in the United States, Version 7.0. In L. M. Vasilas, G. W. Hurt, and C. V. Noble (Eds.). USDA, NRCS, in Cooperation with the National Technical Committee for Hydric Soils. www.nrcs.usda.gov/Internet/FSE_DOCUMENTS/stelprdb1046970.pdf

Veneman, P. L. M. and R. W. Tiner. 1990. *Soil–Vegetation Correlations in the Connecticut River Floodplain of Western Massachusetts*. U.S. Fish and Wildlife Service, Biological Report 90(6). U.S. Government Printing Office, Washington, DC, 51pp.

Vepraskas, M. J., S. J. Teets, J. L. Richardson, and J. P. Tandarich. 1995. *Development of Redoximorphic Features in Constructed Wetlands*. Technical Paper No. 5. Wetlands Research Inc., Chicago, 12p.

Walter, R. C. and D. J. Merritts. 2008. Natural streams and the legacy of water-powered mills. *Science* 319: 299–304.

Ward, J. V., K. Tockner, D. B. Arscott, and C. Claret. 2002. Riverine landscape diversity. *Freshwater Biol.* 47: 517–539.

Wolman, M. G. and L. B. Leopold. 1957. *River Floodplains: Some Observations on Their Formation*. USGS Professional Paper 282-C, U.S. Government Printing Office, Washington, DC.

13

Soils of Tidal Wetlands

Martin C. Rabenhorst and Brian A. Needelman

CONTENTS

Introduction

Within the hydrogeomorphic framework for classifying wetlands and understanding their functional processes, wetlands are described and grouped according to their geomorphology, water source, and hydrodynamics. By geomorphology, we mean the larger landscape and watershed setting within which the wetland occurs. The geomorphology of wetlands is largely responsible for the focusing of surface water or groundwater so as to maintain saturation, flooding, or ponding for significant periods of time. Typical examples are upland depressions or floodplains along riverine systems. Alternatively, in the case of tidal wetlands, the geomorphological setting places the wetlands at an elevation and location in close proximity to a significant tidal water body, such as an estuary or lagoon.

The source of water in a wetland can have a number of dramatic ramifications on soil–water processes, including water chemistry and energy vectors associated with water movement. Direct infall of precipitation can be important in all wetlands, particularly in humid regions. Generally, rainfall or snow melt will be lower in solutes than most surface or groundwater, and will move directly into the soil toward the groundwater, unless slow infiltration causes it to move laterally over the soil surface or in the shallow subsurface zone. Groundwater discharge to wetlands is the dominant source of water for many depressional systems, but can also be an important water source within smaller discharge stream systems (Chapters 3 and 9). While there are a few instances where surface waters may dominate depressional wetland systems, such as the surface-focused recharge wetlands in the prairie pothole region (Chapter 3), surface water is the main source of water entering wetlands associated with tidal flooding conditions. This may occur only occasionally or seasonally within riverine systems, but it occurs regularly in tidal systems.

Hydrodynamics refers to the motion of water and the capacity of the water to accomplish work, such as the transport of sediments, the flushing of hypersaline water, or the transport of nutrients to roots. There is both an energy and a direction associated with water entering and moving through wetlands. The kinetic energy of the water is related to its velocity and may be reflected in the particle size distribution of the suspended load. The direction of surface water has sometimes been classified as vertical (often corresponding to depressional systems), unidirectional (often corresponding to riverine systems), and bidirectional (often corresponding to tidal systems).

Geomorphology of Tidal Wetlands

Geomorphic Models

On the basis of studies in the Chesapeake Bay estuary, Darmody and Foss (1979) described three basic geomorphic types of tidal marshes. Estuarine marshes form in alluvial sediments deposited along tidally influenced rivers and streams (Figure 13.1a). The sediments generally have been eroded from within higher portions of the watershed during storm events and transported downstream to the tidal portion of the stream. During periods of especially high tides, the sediment-borne waters move beyond the channel and over the marsh where the velocity is decreased by the marsh vegetation. The carrying capacity consequently decreases, resulting in sediment deposition. Owing to the relatively low velocity of estuarine streams and rivers, the mineral component in these soils is mostly silts and clays. The mineral content of the marsh soils is dependent on the balance between the magnitude of erosion and deposition of mineral soil from upstream, and the rate of organic matter (OM) production within the marsh. For instance, if a stream provides a significant sediment source in close proximity to the marsh, these soils typically have a higher mineral content and are classified as Entisols. If extended periods occur without significant mineral deposition, organic lenses can be found stratified within the C horizons, or if these periods are especially prolonged (over decades), organic (O) horizons may form. Because these sediments have accumulated under water and have little opportunity for consolidation through dewatering, drying, or compaction, they typically have a high water content and low bulk density. Therefore, these soils tend to have a low-bearing strength and have moderate or high fluidity (Schoeneberger et al. 2012) with an *n*-value of >1 (Soil Survey Staff 1999).

In submerging (transgressive) coastal landscapes, Darmody and Foss (1979) described coastal-type soils forming in marshes behind barrier island systems (Figure 13.1c). These marshes form from organic and mineral sediments within a protected lagoonal setting behind a barrier island. Initially, unvegetated intertidal flats may become colonized by

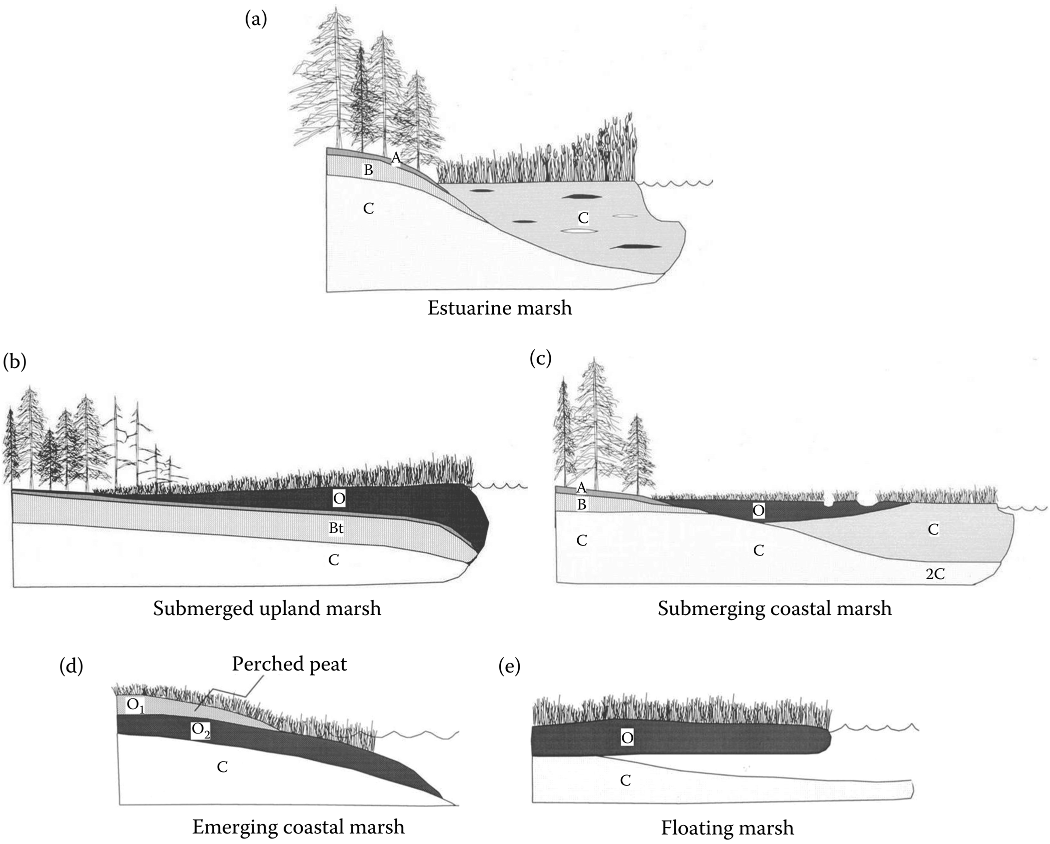


FIGURE 13.1
Tidal marsh types classified according to geomorphologic settings. (a) Estuarine; (b) submerging coastal; (c) submerged upland; (d) emerging coastal; and (e) floating. (Adapted from Darmody, R. G. and J. E. Foss. 1979. *Soil Sci. Soc. Am. J.* 43: 534–541; Stevenson, J. C., M. S. Kearney, and E. C. Pendleton. 1985. *Mar. Geol.* 67: 213–235.)

marsh plants, sometimes aided by the growth of algae (Steers 1977). The growth of the plants provides organic materials directly to the soil and also aids in the trapping of suspended sediment from the tidal waters. Where these marshes occur directly adjacent to a barrier island, the main source of mineral sediment to these soils may be the sandy sediments of the island itself. Therefore, sandy lenses are common within the O horizons, and occasionally, the marsh surface may become buried by a significant deposit of sand during violent storms. Because marshes in these locations are not typically in close proximity to sediment-laden estuarine streams, there is less opportunity for the accumulation of finer-textured mineral components. However, on the landward side of the bay or lagoon, the mineral sediments may be finer in texture, resulting in silty or clayey lenses within organic horizons, or even the formation of fine-textured mineral soils.

The third geomorphic setting for marsh soils in estuaries such as Chesapeake Bay has been called submerged uplands (Darmody and Foss 1979) (Figure 13.1b). The essentially continuous (although punctuated) rise in sea level over the last several thousand years has caused the formation of marsh soils overlying what were once better-drained upland soils on very gently sloping to nearly level landscapes. The O horizons are thinnest at the upland margin of the marsh and usually thicken toward the estuary. Slow rates of sea-level rise of less than a few millimeters per year (actually an apparent sea-level rise caused by both the rising sea level and coastal subsidence) have permitted the vertical accretion of O horizons at rates that keep pace with the sea-level rise (Rabenhorst 1997). The properties of submerged upland-type soils include both those inherited from the former upland soils (such as Bt horizons with high bulk densities) and also those acquired during the formation of organic horizons. Near the margins where the O horizons are thin, the soil classification is strongly affected by the old upland mineral soils. Where the O horizons are thick enough (>40 cm), the soils are classified as Histosols. Also, the salinity and base saturation of the mineral portions of the old submerged soils are typically elevated due to the influence of more saline estuarine waters. Within these settings, there is also a pronounced, although gradual, change in vegetation, with marsh species pioneering as an understory below trees dying from excessive wetness and salinity.

In describing the geomorphic settings of tidal wetlands, Stevenson et al. (1986) added two cases to those of Darmody and Foss (1979) to address two less-common situations encountered in estuaries. In certain regions (such as along the California coastline and in Finland), tectonic activity has been causing an emergence of the coast relative to sea level (regression). Under these circumstances, organic soil horizons that formed at or near sea level have been raised to higher elevations and effectively perched above their zone of formation. These have been termed emerging coastal types (Figure 13.1d) by Stevenson et al. (1986), which they contrast with the submerging coastal type of Darmody and Foss (1979). Occasionally in tidal systems, but more common in lacustrine settings, densely interwoven organic horizons may be underlain by water. Since the marsh soil is effectively buoyed up by entrapped air and the low density of OM, these types have been termed floating marshes (Figure 13.1e).

Geomorphic Processes

Marshes have been differentiated into zones based on elevation and the resulting frequency of tidal inundation, and described as low, middle, or high marshes (Redfield 1972) (Figure 13.2). Low marsh areas are inundated frequently and have also been termed submergence marshes, while high marsh areas are inundated less frequently and have

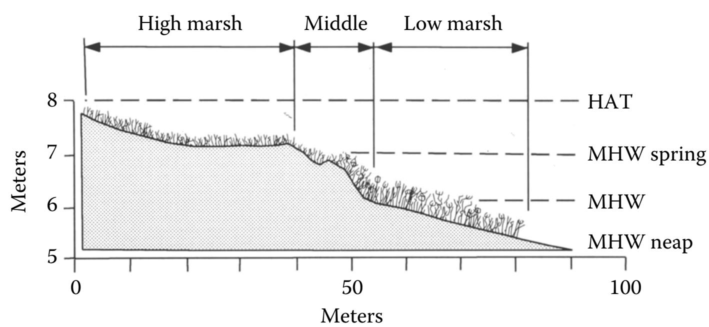


FIGURE 13.2
Classification of marsh zones into low, middle, and high marsh areas based on elevation and frequency of tidal inundation. The mean high water (MHW) for spring and neap tides is shown, as well as the high astronomical tide (HAT). (Adapted from Long, S. P. and C. F. Mason. 1983. *Saltmarsh Ecology*. Blackie Publishing, Glasgow and London, 160pp.)

been termed by some as emergence marshes (Ranwell 1972; Adam 1990). In England (Long and Mason 1983) and in the northeastern United States (Redfield 1972), where tidal ranges are moderate (1–3 m), particular types of vegetation have been reported to be associated with these elevational zones. In the middle portion of Chesapeake Bay estuary where the tidal range is lower (<1 m), the vegetational zonation is less pronounced, although some zonation may still be observed, where, for example, at the highest elevations within some marshes, Spartina patens tends to dominate (Darmody 1975). More detailed investigations have suggested, however, that even if vegetational–elevational associations are observed within an estuary, the patterns may not be applicable to other estuaries (Adam 1990).

In general, both the nature and the geographic proximity of sediment sources have large effects on the mineral components of marsh soils, but the mechanics of physical transport tie the accumulation of mineral sediments in marsh soils both to geomorphology and to hydrodynamics. Because sand grains are larger and require greater transport energy than silt and clay, sandy sediments are usually transported to only relatively short distances in an estuary and are added to marsh soils that are in close proximity to the source. Coastal marshes behind barrier islands commonly receive sandy sediments that are either blown or washed in during storm events. Finer-textured sediments (silt and clay) are more easily transported within the estuary and may be deposited in marshes at greater distances from their origin.

Where marshes occur adjacent to streams and rivers that carry a significant load of eroded sediment from further up the watershed, they will generally receive greater mineral additions and will most commonly be mineral soils (Hydraquents or Sulfaquents). Where the marshes are located farther away from the source of mineral sediment (such as in expansive submerged upland marshes), the relative input of mineral sediments will be less than from those sites nearer the source. These soils will be composed of a greater proportion of organic sediment derived from the marsh vegetation itself, and the soils in these marshes will more likely be organic (Histosols).

Where the mineral load to marshes is significant (i.e., where marsh growth and accretion are strongly associated with mineral deposition [low OM soils]), the elevation of the marsh surface may have an important effect on sedimentation rates (Steers 1977). Because areas of lower elevation are submerged more frequently, they have greater potential for receiving sediment than those at higher elevations that are submerged less frequently. However, portions of the marsh directly adjacent to the sediment-laden stream may be rapidly accreting levees at slightly higher elevation because they receive greater sediment deposition as they decrease the velocity of the flooding water, lowering its energy and its transport capacity. Because the density and structure of marsh plants affect the movement of floodwater across the marsh, plant type and vigor can also affect the accumulation of mineral sediments.

Human alterations have been extensive in many tidal wetland systems, often fundamentally changing geomorphological conditions. Water levels and sediment supply are altered through ditching, diking, and impoundments. Upstream alterations such as dams and land management practices also affect tidal wetland geomorphology. The general decline of sediment delivery to coastal zones due to dams and improved sediment control practices represents a threat to marshes dependent on sediment delivery to accrete sufficiently to keep pace with sea-level rise (Weston 2014). Restoration practices are often most effective when they restore natural, hydrological, and other geomorphological conditions, although this should be done in the context of watershed-scale dynamics, landscape evolution, and climate change scenarios (Crooks 2012).

Factors Affecting Marsh Vegetation Production and Community Composition

A great deal has been written describing vegetational succession in tidal marshes, and the breadth and extent of the topic are beyond the scope of this chapter. Nevertheless, a cursory sketch of the factors that govern the development, succession, and distribution of marsh vegetation is warranted. For a given marsh, the combination of elevation and tidal characteristics defines the frequency and duration of tidal inundation. Marsh plants are variably adapted to survival under submerged conditions, which helps to determine their distribution along elevational gradients. Also contributing to the distribution of plant populations is the salinity of estuarine and marsh soil pore waters. Salinity gradients are commonly observed within tidal estuaries between fresher portions toward the headwaters and more saline portions nearer the ocean. Thus, on a broad scale, the general vegetational distribution within an estuary will in part be related to the salinity in that portion of the estuary. Locally, salinity tolerance may affect plant distribution within a marsh, while elevation and frequency of tidal inundation may also affect the salinity of marsh soils. As a contributing factor, the physical nature of the mineral substrate is probably less influential on marsh vegetation than salinity or the tidal regime, but it can vary greatly between highly sandy sediments with very little fines to sediments dominated by silts and clays and essentially no sand.

Plant production, particularly belowground, is a dominant factor controlling the contribution of soil volume changes (accretion) to the overall marsh elevation changes in tidal wetlands, particularly in wetlands without high sediment inputs (Nyman et al. 2006). Marsh plants tend to exhibit an optimal (maximum) growth rate at moderate inundation levels, with decreased growth associated with either increased or decreased inundation (Morris et al. 2002). Marshes are most resilient to sea-level rise when they are situated below this optimum relative to sea level, such that productivity will increase as sea level rises, allowing for greater accretion rates. Marshes situated above their optimum relative to sea level will respond to increased sea levels with decreased productivity, causing decreased accretion likely leading to degradation and eventual submergence. The elevational growth range of plant species decreases with decreasing tidal range, such that wetlands with small tidal ranges are most susceptible to degradation due to accelerated sea-level rise (McKee and Patrick 1988). The magnitude of sea-level rise to which a wetland is resilient prior to undergoing degradation has been termed elevation capital (Cahoon and Guntenspergen 2010).

Water Source in Tidal Wetlands

Waters that enter tidal wetlands can come from any of the three possible sources, including precipitation, surface water, or groundwater, although in comparison to the other two, groundwater inputs to tidal marshes are probably small. Surface waters represent the dominant water source for most tidal wetlands. Both astronomical and weather-related (storm) tides cause waters to flood and submerge the marsh surface at varying frequencies, depending on factors such as elevation and location. Of particular significance to tidal wetlands is the nature of these tidal floodwaters. The chemical composition of tidal waters ranges widely depending on their proximity to ocean water. Generally, coastal and estuarine marshes have salinities and levels of ionic solutes significantly above those of freshwater, and in some cases, they may be orders of magnitude greater, eventually approaching levels found in seawater. Some tidally influenced marshes located along the upper reaches of estuarine rivers or in the interiors of coastal deltas may, however, have levels of soluble constituents similar to those of freshwater systems (Baxter 1973; Miehlich 1986). Elevated levels of solutes in tidal waters include many required plant nutrients, such as K, Ca, Mg, and S. Soluble N and P levels, however, tend to be fairly low in ocean waters relative to fresh waters (Long and Mason 1983). Therefore, much of the N and P that enters tidal marsh systems comes from either soluble or absorbed nutrients transported from upland sources. In addition, some N-fixing microorganisms (free living or symbiotic) occur in marsh soils and contribute to plant-available N.

By comparison, meteoric waters that enter tidal marshes contain very low levels of solutes. The significance of meteoric water entering a tidal marsh is dependent on the frequency with which the marsh is inundated with tidal water. In lower portions of the marsh that are flooded frequently by semidiurnal tides, the dilution affected by meteoric waters may be insignificant. Under these conditions, the marsh pore water is dominated by the chemistry of the tidal water. However, at higher elevations within the marsh, which may undergo extended periods without flooding by tidal water, the intrusion of fresh meteoric water may significantly affect short-term changes in the marsh pore water chemistry.

Hydrodynamics of Tidal Wetlands

Tidal Frequency and Range

In most of the coastal marshes of the world, astronomical conditions cause two tidal cycles per day with a period of approximately 12.5 h. The range in elevation between high and low tides varies widely depending largely on geography. Along the European Atlantic coast, the tidal range is approximately 3 m, and along the North American Atlantic coast, it ranges from 1 to 3 m. The Bay of Fundy is notorious for having the largest tidal range in the world—about 18 m (Steers 1977). The coastal bays of Delaware and Maryland (Sinepuxent, Assawoman) and the Chesapeake Bay estuary typically have smaller tidal ranges of <1 m, as do those along the gulf coast of Louisiana (Chabreck 1972). Perhaps, the lowest tidal range occurs along the Baltic coast of Sweden, where it is 0.3 m (Ranwell 1972).

Superimposed upon the normal ranges of the twice-daily cycles of high and low tides are lunar cycles causing unusually high (spring) high tides when the sun, moon, and earth are aligned (~1.5 days after full and new moons) and unusually low (neap) high tides, that occur 7 days after spring tides (Figure 13.3). While particular patterns may vary from location to location, the magnitude of the tidal range is also related to annual cycles as illustrated in Figure 13.4. In addition to the astronomical conditions affecting heights of high and low tides, which can be calculated and predicted, are unpredictable meteorological conditions such as barometric pressure and prevailing winds, that can also affect tides. These are usually translated into infrequent and irregular storm tides with varying return frequencies. Analysis of a 50-year record of tides from Solomon's Island, MD, indicated that the frequency of occurrence of above-normal high tides was described reasonably well by a log function (Rabenhorst 1997) (Figure 13.5).

Other Factors Affecting Hydrodynamics

While the daily tidal range may average 1 m or more in many areas, the fluctuation in water tables within most marsh soils is much less. The work of Haering (1986)

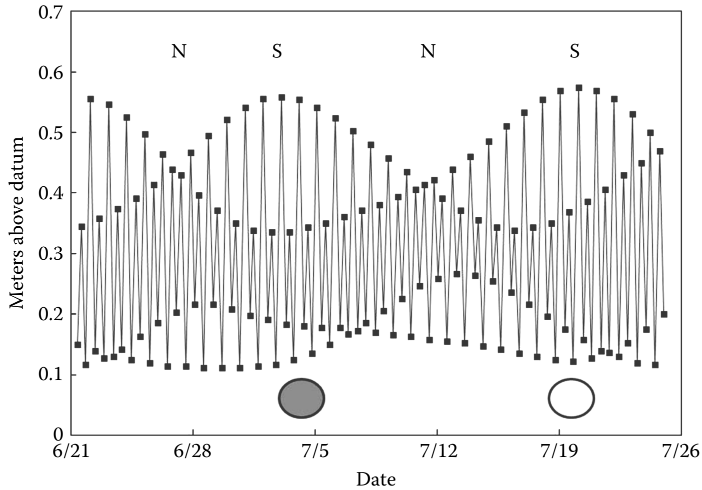


FIGURE 13.3
Simulated high and low tides at Solomon's Island (Chesapeake Bay) during a 5-week period. Superimposed upon the diurnal tides are lunar cycles causing unusually high, high tides (spring tides—S) (July 3 and July 20) near the time of full and new moons, and unusually low, high tides (neap tides—N) that occur approximately midway between spring tides (June 26 and July 11). Note the new moon on July 4 and full moon on July 20.

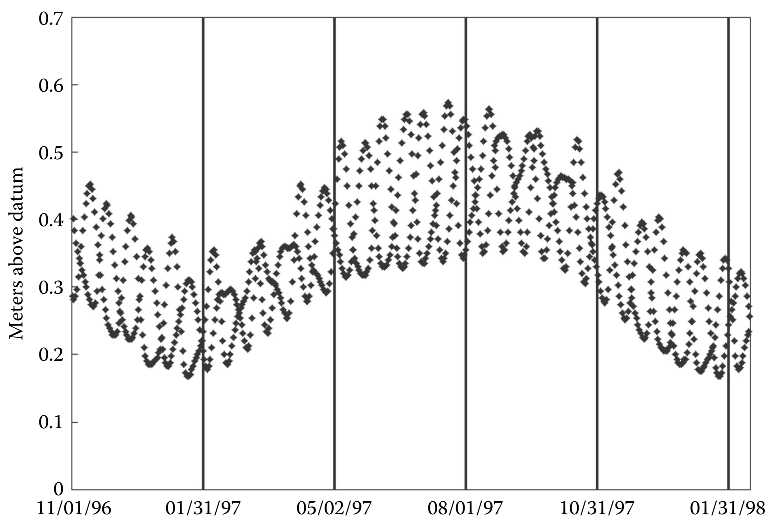


FIGURE 13.4
Simulated high tides at Solomon's Island (Chesapeake Bay) from 10/96 to 2/98. Note the annual cycle for this particular site that has higher spring tides during the summer months and lower spring tides during the winter months.

and Haering et al. (1989) indicated that the rise and fall of marsh soil water tables are only significant in the vicinity nearest the tidal creeks, with measurable effects occurring only within 20 m of the tidal creek. This is apparently due to the periodicity of tides (12.5-h cycle) and the modest rates of hydraulic conductivity in marsh soils. Thus, one should avoid embracing the simplistic caricature of water tables in marsh soils rising and falling with the daily tidal cycles.

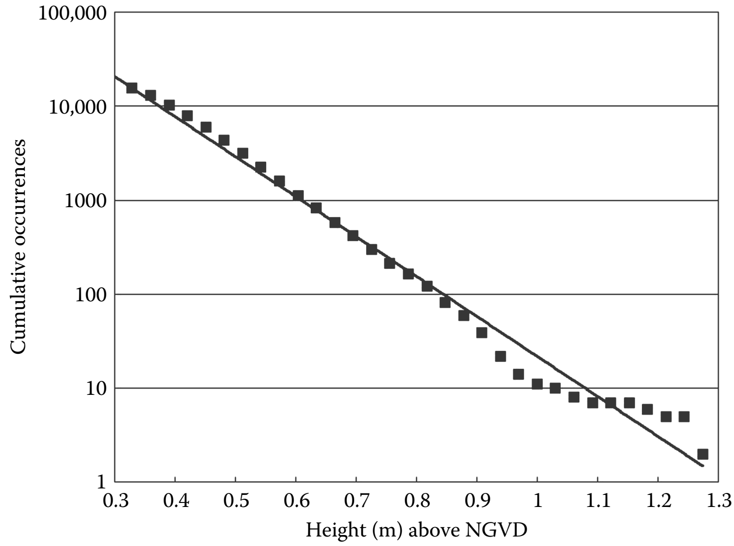


FIGURE 13.5
Analysis of a 50-year record of tidal measurements from Solomon's Island, MD, indicates that the cumulative frequency of occurrence of above-normal high tides is described well by a log function. The height of tides is given in relation to the national geodetic vertical datum of 1929 (NGVD) (formerly the sea-level datum of 1929.)

At the lowest elevations within the marsh, the soils are generally flooded twice each day by tidal waters, and the water movement is predominantly bidirectional. The result is that the soil salinity is rather constant and very near that of the flooding water. In contrast, at the highest elevations within the marsh, the soils are flooded only by extreme tides and storm tides. Depending on the balance between precipitation and evapotranspiration, the water tables in these soils may drop to varying depths below the surface. During periods of high evapotranspiration, salts may accumulate in the marsh soil, causing highly saline conditions. During periods of rainfall, the salinity of the soils may be lowered considerably by leaching with meteoric water.

Stream Flow

The hydrodynamics in tidal wetlands can be affected by marsh accretion and succession, leading to the development of creeks or streams. Initially, on unvegetated sand or mud flats, tidal waters will basically have a uniform ebb and flow across the area. As vegetation becomes established, flow is restricted and channelized. Eventually, most of the water movement becomes restricted to channelized flow within tidal creeks. Generally speaking, the presence of vegetation promotes accretion in those areas, while at the same time, tidal scour maintains or deepens creeks that can produce a steep-sided channel (Long and Mason 1983).

Most tidal marshes are interlaced with a network of tidal streams, often following a dendritic, but sometimes a trellis or rectangular, pattern (Steers 1977). Most tidal streams reach the bank full roughly 360 times per year in the mid-marsh section, which stands in contrast to most alluvial systems, that reach the bank full only occasionally (a few times per year at most). The migration of stream channels within the marsh can be very dynamic. Relative to other alluvial systems, the meanders move rapidly at rates reported to be up to 100 m per century (Long and Mason 1983). Most of the water in tidal marshes enters and leaves via the system of tidal creeks. Therefore, the marsh soil pore water near the creek banks is most similar to the tidal water itself, but its chemical composition changes dramatically with distance from the creek bank. Usually, only during spring tides or storm tides will the water overflow the banks and cause the marsh to become fully submerged.

The flow regimes of streams in tidal marshes are distinctly bidirectional, with water moving against the bed gradient during flow regimes and moving along the bed gradient during ebb conditions. Some have suggested that the velocity and energy of water in tidal streams is lower during flow and greater during ebb conditions because of the effects of gravity (Long and Mason 1983). That there may be some variation in flow and ebb conditions is unquestionable, although the nature of this hysteresis may not be predictable. The relative velocity and energy of water flowing in tidal creeks during periods of flow and ebb play important roles in determining the relative balance between accretion/sedimentation and erosion in marsh development.

Although the water flow in tidal marshes is mainly bidirectional, there may be particular conditions when vertical and unidirectional flow may also occur. At higher elevations in the marsh that are only occasionally flooded by tidal waters, the reception and infiltration of meteoric water results in vertical movement and a downward-leaching vector in the uppermost soil horizons above the free water surface. In the lower portions of the marsh, that are frequently inundated by tidal waters, meteoric infall would not have any noticeable effect. Also, storm events in the watershed supplying flow to an estuarine stream may have the effect of causing flood conditions in parts of the estuarine marsh that may resemble the unidirectional flow typical of other flooded alluvial systems.

Geochemistry of Tidal Wetlands

Effects of Peraquic Conditions

While the water tables in some of the higher portions of tidal wetlands may occasionally drop during periods of high evapotranspiration between spring tides, the water source in these wetlands is more or less constant, and the wetland hydrology is generally maintained. Because the hydrology is not particularly dependent on seasonal conditions and variations, the high water tables in tidal wetlands are essentially permanent, leading to what has been described as "peraquic" conditions in the soil (Soil Survey Staff 1999). Also for a variety of reasons (including dispersion and low permeability of estuarine sediments, and low hydrostatic head within tidal wetlands), the rate of water movement through wetland sediments is slow (Knott et al. 1987). Together, these factors result in wetland soils that not only have very low electrochemical redox potentials, but also maintain these low redox potentials throughout much of the year (Chapter 4). Because the development of reducing conditions in these wetland soils is microbially mediated, and because microbial activity is temperature dependent, there can still be significant seasonal trends in the soil redox conditions that may be related to soil temperature and to availability of labile C sources during plant senescence (Feijtel et al. 1988; Oenema 1988; Krairapanond et al. 1991).

Sulfidization and Methanogenesis

As heterotrophic bacteria decompose organic materials, they utilize various compounds and ions as electron acceptors under various Eh regimes and proceed to lower the redox potential as one acceptor is depleted and another is utilized. Generally, they are utilized in the order of O_2, NO_3^-, Mn^{4+}, Fe^{3+}, SO_4^{2-}, and CO_2 (Chapter 4). The peraquic conditions of tidal marsh soils create two important conditions that influence soil development. First, OM accumulates from the primary marsh vegetation that provides an energy source for microbes. Second, diffusion of atmospheric oxygen into the saturated soils is inhibited. Therefore, dissolved oxygen is quickly depleted, and the microbes move on to alternate electron acceptors. In tidal marshes, it is common for soils to become sufficiently reduced so that microbes utilize sulfate as the primary electron acceptor. Therefore, tidal marsh soils often present a combination of conditions that are optimal for sulfidization, including OM as a microbial energy source, low redox potentials, the presence of sulfate as an electron acceptor, and sulfate-reducing bacteria (Rabenhorst and James 1992). While Goldhaber and Kaplan (1982) have indicated that rates of sulfate reduction are independent of sulfate concentrations when >10 mM, the work of Haering (1986) suggests that sulfur accumulation in tidal marsh soils may begin to be limited by sulfate concentrations only when levels in the estuarine water drop below 1 mM. This would suggest that most estuarine waters with salinities greater than 1 or 2 ppt generally will have adequate sulfate for sulfate reduction during microbial oxidation of organic carbon (Equation 13.1).

$$SO_4^{2-} + 2CH_2O \xrightarrow{SO_4^{2-}\text{ reducing bacteria}} H_2S + 2HCO_3^- \quad (13.1)$$

If a source of reactive Fe is present in the marsh soil where sulfate reduction is occurring, iron sulfide minerals will form (Rabenhorst and James 1992). Both monosulfide (Equation 13.2) and disulfide (Equations 13.3 and 13.4) species can form (Rabenhorst 1990), although the disulfide (pyrite—FeS_2) is the thermodynamically favored phase. Both individual crystals of pyrite and framboids will form in marsh soils. It has also been demonstrated that soluble sulfide can chemically reduce iron oxides and result in the formation of solid-phase iron sulfide minerals as shown in Equation 13.5 (Rabenhorst 1990; Rabenhorst et al. 2010; Fanning et al. 2012).

Both Griffin and Rabenhorst (1989) and Rabenhorst and James (1992) have discussed the ways in which various factors necessary for sulfidization could affect or limit the formation of sulfides in tidal marsh systems. Any of a number of heavy and transition metals could substitute in small quantities for Fe in the pyrite structures. Thus, one of the environmental implications of sulfidization in tidal marsh soils is that heavy metals may accumulate as sulfides and other phases in the soils and sediments of tidal marshes and help ameliorate contaminated estuaries (Lindau and Hossner 1982; Griffin et al. 1989).

$$Fe^{2+} + S^{2-} \rightarrow FeS \quad (13.2)$$

$$FeS + S^o \rightarrow FeS_{2(\text{pyrite})}(\text{simplified}) \quad (13.3)$$

or

$$Fe^{2+} + S_x^{Y-} \rightarrow FeS_2 + (X-2)S^{(Y-2)-} \quad (13.4)$$

$$2FeOOH + 3H_{2S} \rightarrow FeS_2 + FeS + 4H_2O \quad (13.5)$$

Because metal sulfides can accumulate in tidal marsh soils, they possess the potential for acid sulfate weathering if disturbed through dredging or excavating operations. When soil materials containing sulfide minerals are exposed to aerobic conditions, the S and Fe in sulfide minerals can be oxidized through microbial activity (followed by hydrolysis of the Fe[III]), generating acidity in the form of sulfuric acid (Equation 13.6). If the neutralizing potential (through $CaCO_3$) or the buffering capacity of the soil is not adequate to counteract the acid generated, the soil itself can become extremely acid. In this way, soil materials from some tidal marshes, that under natural (peraquic) conditions usually have pH values that are neutral or slightly alkaline, can develop pH values as low as 3 or less when disturbed and oxidized.

$$FeS_2 + 5/2H_2O \rightarrow FeOOH + 2H_2SO_4 \tag{13.6}$$

The final electron acceptor that may be used by heterotrophic bacteria is CO_2, in which CO_2 is reduced to CH_4 in a methanogenic pathway. Methanogenesis occurs in most peraquic tidal marsh soils, but is generally only a significant component when sulfate concentrations are depleted because sulfate reduction is favored over methanogenesis (Widdell 1988). Methane emissions are generally very low below a threshold of approximately 18 ppt salinity (polyhaline), assuming that there is no source of sulfate unassociated with salinity (Poffenbarger et al. 2011) (Figure 13.6). Mesohaline marsh soils (5–18 ppt salinity) have been found to have moderate but significant methane emission rates. Fresh and oligohaline marsh soils (0–5 ppt salinity) have a wide range of methane emissions; the low methane emissions observed from some of these soils are likely due to a lack of peraquic conditions.

There is increasing interest in tidal wetland soils for the mitigation of rising greenhouse gas concentrations due to their high rates of carbon sequestration (Chmura et al. 2003; Needelman and Hawkes 2012; Kennedy et al. 2014). Methodologies are being developed

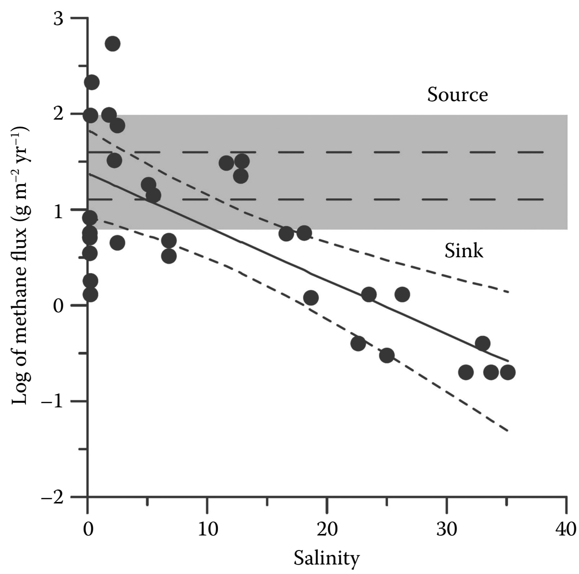


FIGURE 13.6
Tidal marsh methane emissions versus salinity. (From Poffenbarger, H., B. A. Needelman, and J. P. Megonigal. 2011. *Wetlands* 31: 831–842). The curve is a linear fit of salinity against log-transformed methane flux data with 95% confidence intervals (pointwise). The horizontal gray band represents the methane emission equivalents of the 5% and 95% quantiles of tidal marsh carbon sequestration rates reported by Chmura et al. (2003); the horizontal dashed lines are the 25% and 75% quantiles of this data set (methane equivalence based on a global-warming potential of 25.)

for voluntary carbon-crediting programs, such as the Verified Carbon Standard, such that restoration and conservation projects may be eligible to receive carbon financing.

Both methane and nitrous oxide are potent greenhouse gases that can be emitted from tidal wetlands, potentially offsetting the carbon sequestration benefit. A significant challenge for greenhouse gas accounting is the estimation of methane emissions from fresh and brackish wetlands. The low redox potentials found in tidal marsh soils should render nitrate highly unstable such that it would either be denitrified and lost as N_2 or reduced to ammonium (Chapter 4). Nitrous oxide emissions will generally be lowered due to the restoration of tidal wetlands due to maintained or increased water levels, but restoration practices that lower water tables may require nitrous oxide accounting (such as impoundment breaching and wetland creation using dredged material).

Morphology and Classification of Tidal Marsh Soils

Morphology and Horizonation

The dominant soil horizons described in marsh soils are O, A, and C. Only occasionally are B horizons described, and usually, their properties are inherited from some prior episode of pedogenesis. Whether a marsh soil horizon is designated O, A, or C largely depends on the relative proportion of mineral and organic components. Where marsh vegetation is actively growing in areas distant from a sediment source, organic (O) horizons will form, whereas if conditions favor rapid accumulation of mineral sediments relative to organic materials, C or A horizons will form. It is also common during marsh soil genesis for the balance between these conditions to change so that occasional organic horizons or lenses may be interspersed within a dominantly mineral soil and vice versa.

Most organic soil horizons are dark in color with Munsell value/chroma of 3/2 or darker. They are differentiated by the degree of decomposition of the organic materials, that is mainly based on the quantity of recognizable plant materials remaining after rubbing (such as Oi, Oe, and Oa, as the degree of decomposition increases) (see Chapters 1 and 6).

The mineral soil horizons are almost always gleyed or gray in color with a Munsell chroma of 2 or less, and often 1 or less, regardless of whether they are sandy, loamy, or clayey. These gray colors can be attributed to the colors of the mineral grains, that lack coatings of iron oxides more typical of upland soils. When the mineral sediments are sandy, they tend to have a higher density and lower *n*-value (nonfluid). The mineral horizons in tidal marshes that are loamy or clayey, however, tend to have a lower density, high water content, and consequently a high fluidity, sometimes designated as a high *n*-value ($n > 1$), which is one of the common characteristics of tidal marsh soils. The notable exception to this is when loamy subsoils of submerged upland marshes underlie more recently accumulated organic or mineral horizons. In these cases, the subsoil horizons are B rather than C horizons, and have properties inherited from upland pedogenic processes. These horizons have bulk densities typical for upland soils (1.3–1.7 g/cm^3), are nonfluid (low *n*-value; $n < 0.7$), and may even have inherited features such as illuvial clay films. It is also common for these soil B horizons to contain soft masses and concentrations of Fe (redox concentrations) within a gleyed or depleted matrix, that are generally absent from tidal marsh soils. Adaptations of some marsh plants enable them to oxygenate the environment immediately adjacent to their roots, forming an oxidized rhizosphere. If soluble ferrous Fe is present, the Fe may precipitate as oxyhydroxides, forming redox concentrations in the form of soil pore linings or coatings on roots.

Classification

If as much as 40 cm of the upper 80 cm of a tidal marsh soil is composed of organic soil materials, it is classified as a Histosol. Most other tidal marsh soils are classified as Entisols. Histosols are divided into the suborders Fibrists, Hemists, and Saprists, depending on the degree of decomposition in the subsurface organic horizons. In general, organic horizons forming in tidal marshes of cooler regions will be less decomposed than those forming in warmer regions. It is for this reason that Fibrists are mapped mainly in the tidal marshes of cold areas such as Alaska. While both Hemists and Saprists can be found in tidal marshes all along the Atlantic, Gulf, and Pacific coasts, Hemists are dominant in New England and the Pacific Northwest, while Saprists dominate in the southeast Atlantic and Gulf coastal regions. At another level of classification, Histosols are differentiated between Terric and Typic subgroups based on whether they have a mineral layer 30 cm or more thick that has its upper boundary between 40 and 130 cm. A simpler, though not quite so accurate, way to describe this is to say that those soils where the base of the organic horizons is shallower than 130 cm are Terric, while those that have deeper organic materials are Typic. Some organic soils in tidal marshes of northern New England (Breeding et al. 1974) overlie rock at a shallow depth, which causes them to be classified as Lithic subgroups. Similarly, Histosols in Florida overlying coral and limestone are also classified as Lithic subgroups.

A significant characteristic used in classifying both organic and mineral tidal marsh soils (at the great group level) is the presence or absence of sulfidic materials within 100 cm of the soil surface (within 100 cm of the surface for Histosols and within 50 cm for Entisols). Sulfidic materials must contain a sufficient quantity of sulfide minerals (such as pyrite) so that when incubated under moist and ambient conditions, "the pH decreases by 0.5 or more units to a value of 4.0 or less." Thus, the sulfide is sufficient for evidencing the potential for acid sulfate weathering (Soil Survey Staff 2010, p. 29).

Tidal marsh soils, that are dominated by mineral soil materials, are classified in the suborder of Aquents (Entisols). They are further differentiated based on the presence or absence of sulfidic materials and the nature and characteristics of the mineral horizons. If the soils contain sulfidic materials, they are classified as Sulfaquents. However, if sulfidic materials are absent and the soils are loamy and have a high *n*-value, they are classified as Hydraquents, while those that are predominantly sandy are classified as Psammaquents.

In submerged upland tidal marshes, if the recently accreted organic sediments are less than 40 cm thick, the soils will be classified based on the nature of the submerged soil. These submerged soils contain properties both inherited from the previous pedogenic environment and properties acquired from the present marsh environment. Gardner et al. (1992) have reported marshes forming over Spodosols on the South Carolina coastal plain, and soils with argillic (Bt) horizons have been reported under marshes in Chesapeake Bay (Stolt and Rabenhorst 1991; Rabenhorst 1997). In Dorchester County, MD, the Sunken soil series was established to accommodate tidal marsh soils with thin organic horizons (<20 cm) overlying soils that were previously Aquults (Brewer et al. 1998). The effect of brackish water diffusing into these soils following tidal submergence has resulted in their being changed into Alfisols, and they are classified as Endoaqualfs (Figure 13.7). Because of the alluvial nature of tidal marsh soils, it is not uncommon for organic horizons or lenses to be interspersed within a mineral soil. In some classes of Aquents, where

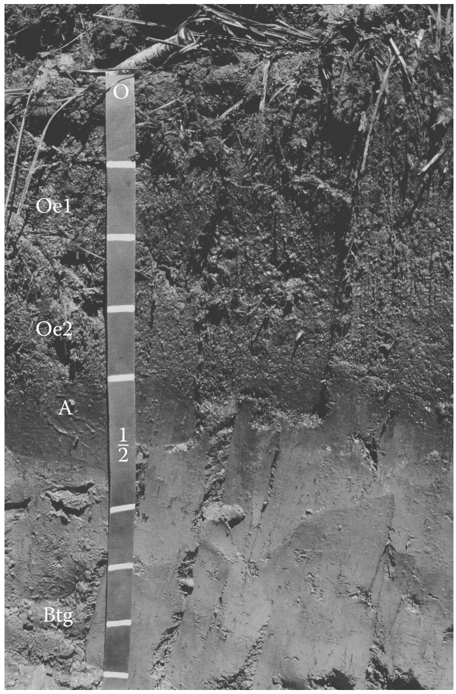


FIGURE 13.7
Profile of a soil in the Honga series (loamy, mixed, euic, mesic, and Terric Sulfihemists) from Dorchester County, MD. This soil was probably a Typic Endoaquult before a gradual rise in sea level caused the accumulation of organic materials at the surface, that are now greater than 40 cm in thickness. The mineral subsoil contains a relict argillic (Btg) horizon that formed under a different pedogenic regime.

a buried organic horizon at least 20 cm thick beginning within the upper meter of the soil is recognized, it is accommodated in a Thapto–Histic subgroup.

As a result of the work by Demas (1998); Demas and Rabenhorst (1999, 2001); Demas et al. (1996); and changes to the definition of soil in the second edition of *Soil Taxonomy* (Soil Survey Staff 1999), the substrata of shallow subtidal wetlands (characterized by submerged aquatic vegetation, rather than emergent vegetation) have been recognized as soils.

Identification of Hydric Soils in Tidal Wetlands

The identification and delineation of hydric soils in tidal wetlands is probably among the easiest determinations to make for several reasons. First, they are geomorphologically constrained to locations essentially at sea level. Second, because of the nature of the tidal hydrology, the water table is basically permanent. Thus, unlike many hydric soils that have seasonally high water tables that drop significantly during certain times of the year, soil water tables in tidal wetlands can be readily observed at or near the soil surface any time of the year. Because these soils have peraquic moisture regimes, and because tidal water is often brackish or saline, there is commonly a distinctive vegetation community of obligate hydrophytes or halophytes that occupy the hydric soils of tidal wetlands. While none of these characteristics is a soil morphological characteristic, they are nevertheless diagnostic for wetland identification. In addition, however, there are numerous soil morphological features that indicate the presence of hydric soils.

As was discussed above, many tidal marsh soils are Histosols, which (by itself) is a diagnostic field indicator (field indicator A1; USDA-Natural Resources Conservation Service 2010) in all land resource regions (LRRs) (Chapters 8 and 9). In addition, the presence of a histic epipedon (alone) is an accepted field indicator that the soil is hydric (field indicators A2 and A3). Although not necessarily applicable in all LRRs, there are also several field indicators tied to the occurrence of relatively thin layers of organic soil materials (muck, mucky peat, or peat) at the soil surface (field indicators A8, A9, A10, S2, and S3). For any brackish or saline tidal marsh, the presence of the aroma of hydrogen sulfide gas not only indicates that the necessary conditions for sulfidization were met, but also that the soil is considered to be hydric (field indicator A4). In the case of Hydraquents or Psammaquents, in tidal marshes that lack histic epipedons or sulfidic materials, all of them will nearly meet one or more of the indicators related to low chroma matrix colors (gleyed or depleted) in the upper portion of the soil (field indicators S4, F2, and F3). The soils of tidal marshes are generally so clearly hydric that a given soil will often demonstrate numerous field indicators.

In some areas with very gently sloping landscapes, such as in areas of submerged upland marshes, there may be transitional zones grading from tidal wetlands to nontidal wetlands. There may be areas of nontidal hydric soils within a meter or so of mean high water that are only occasionally inundated by storm tides (Rabenhorst 1997). These soils would need to be identified based on the field indicators used in the general vicinity, that are applicable to nontidal wetlands. Indicator F20 (Anomalous Bright Loamy Soils) was specifically developed for nontidal wetlands in close proximity to tidal areas.

References

Adam, P. 1990. *Saltmarsh Ecology*. Cambridge University Press, Cambridge, 461pp.

Baxter, J. 1973. *Morphological, Physical, Chemical, and Mineralogical Characteristics of Some Tidal Marsh Soils in the Patuxent Estuary*. MS thesis, Department of Agronomy, University of Maryland, College Park.

Breeding, C. H. J., F. D. Richardson, and S. A. L. Pilgrim. 1974. *Soil Survey of New Hampshire Tidal Marshes*. New Hampshire Agricultural Experiment Station and University of New Hampshire, and USDA-SCS. Research Report No. 40. Durham, NH.

Brewer, J. E., G. P. Demas, and D. Holbrook. 1998. *Soil Survey of Dorchester County, Maryland*. USDA-NRCS. In Cooperation with Maryland Agricultural Experiment Station, Maryland Department of Agriculture, and Dorchester Soil Conservation District, U.S. Government Printing Office, Washington, DC, 178pp. 31 maps.

Cahoon, D. R. and G. R. Guntenspergen. 2010. Climate change, sea-level rise, and coastal wetlands. *Natl. Wetlands Newsl.* 32: 8–12.

Chabreck, R. H. 1972. *Vegetation, Water and Soil Characteristics of the Louisiana Coastal Region*. Louisiana State University and Agricultural Experiment Station Bulletin 664.

Chmura, G. L., S. C. Anisfeld, D. R. Cahoon, and J. C. Lynch. 2003. Global carbon sequestration in tidal, saline wetland soils. *Glob. Biogeochem. Cycles* 17: 1–22.

Crooks, S. 2012. Planning and design considerations for coastal habitat restoration in the face of climate change. In B. A. Needelman, J. Benoit, S. Bosak, and C. Lyons (Eds.) *Restore–Adapt–Mitigate: Responding to Climate Change through Coastal Habitat Restoration*. Restore America's Estuaries, Washington, DC, pp. 23–32.

Darmody, R. G. 1975. *Reconnaissance Survey of the Tidal Marsh Soils of Maryland*. MS thesis, Department of Agronomy, University of Maryland, College Park.

Darmody, R. G. and J. E. Foss. 1979. Soil–landscape relationships of the tidal marshes of Maryland. *Soil Sci. Soc. Am. J.* 43: 534–541.

Demas, G. P. 1998. *Subaqueous Soil of Sinepuxent Bay, Maryland*. PhD dissertation, University of Maryland, College Park.

Demas, G. P. and M. C. Rabenhorst. 1999. Subaqueous soils: Pedogenesis in a submersed environment. *Soil Sci. Soc. Am. J.* 63: 1250–1257.

Demas, G. P. and M. C. Rabenhorst. 2001. Factors of subaqueous soil formation: A system of quantitative pedology for submersed environments. *Geoderma* 102: 189–204.

Demas, G. P., M. C. Rabenhorst, and J. C. Stevenson. 1996. Subaqueous soils: A pedological approach to the study of shallow-water habitats. *Estuaries* 19: 229–237.

Fanning, D. S., M. C. Rabenhorst, B. R. James, D. P. Wagner, and D. C. Lowery. 2012. Rethinking sulfidization and the role of hydrogen sulfide. Presented at the *7th International Acid Sulfate Conference*. Vaasa, Finland, August 26–September 1, 2012. *Conference Proceedings*, pp. 44–46. http://arkisto.gtk.fi/op/op56/op_056.pdf

Feijtel, T. C., R. D. DeLaune, and W. H. Patrick Jr. 1988. Seasonal pore-water dynamics in marshes of Barataria Basin, Louisiana. *Soil Sci. Soc. Am. J.* 52: 59–67.

Gardner, L. R., B. R. Smith, and W. K. Michener. 1992. Soil evolution along a forest–salt marsh transect under a regime of slowly rising sea level, southeastern United States. *Geoderma* 55: 141–157.

Goldhaber, M. B. and I. R. Kaplan. 1982. Controls and consequences of sulfate reduction rates in recent marine sediments. In J. A. Kittrick, D. S. Fanning, and L. R. Hossner (Eds.) *Acid Sulfate Weathering*. SSSA Special Publication 10, American Society of Agronomy, Madison, WI, pp. 19–36.

Griffin, T. M. and M. C. Rabenhorst. 1989. Processes and rates of pedogenesis in some Maryland tidal marsh soils. *Soil Sci. Soc. Am. J.* 53: 862–870.

Griffin, T. M., M. C. Rabenhorst, and D. S. Fanning. 1989. Iron and trace metals in tidal marsh soils of the Chesapeake Bay. *Soil Sci. Soc. Am. J.* 53: 1010–1019.

Haering, K. C. 1986. *Sulfur Distribution and Partitionment in Chesapeake Bay Tidal Marsh Soils*. MS thesis, University of Maryland, College Park, 172pp.

Haering, K. C., M. C. Rabenhorst, and D. S. Fanning. 1989. Sulfur speciation in Chesapeake Bay tidal marsh soils. *Soil Sci. Soc. Am. J.* 53: 500–505.

Kennedy, H., D. M. Alongi, A. Karim, G. Chen, G. L. Chmura, S. Crooks, J. G. Kairo, B. Liao, G. Lin, and T. G. Troxler. 2014. Coastal wetlands. In T. Hiraishi, T. Krug, K. Tanabe, N. Srivastava, J. Baasansuren, M. Fukuda, and T. G. Troxler (Eds.) *2013 Supplement to the 2006 IPCC Guidelines for National Greenhouse Gas Inventories: Wetlands*. IPCC, Switzerland.

Knott, J. F., W. K. Nuttle, and H. F. Hemond. 1987. Hydrologic parameters of salt marsh peat. *Hydrol. Process.* 1: 211–220.

Krairapanond, N., R. D. DeLaune, and W. H. Patrick Jr. 1991. Seasonal distribution of sulfur fractions in Louisiana salt marsh soil. *Estuaries* 14(1): 17–28.

Lindau, C. W. and L. R. Hossner. 1982. Sediment fractionation of Cu, Ni, Zn, Cr, Mn, and Fe in one experimental and three natural marshes. *J. Environ. Qual.* 11: 540–545.

Long, S. P. and C. F. Mason. 1983. *Saltmarsh Ecology*. Blackie Publishing, Glasgow and London, 160pp.

McKee, K. L. and W. H. Patrick Jr. 1988. The relationship of smooth cordgrass (*Spartina alterniflora*) to tidal datums: A review. *Estuaries* 11: 143–151.

Miehlich, G. 1986. Freshwater-marsh of the Elbe river. In H. P. Blume (Ed.) *Guidebook for a Tour of Landscapes, Soils and Land Use in the Federal Republic of Germany.* 13th Congress International Society of Soil Science, Hamburg, Germany. Mittielungen der Deutschen Bodenkundlichen Gesellschaft. Band 51. ISSN-0343-107X, pp. 99–128.

Morris, J. T., P. V. Sundareshwar, C. T. Nietch, B. Kjerfve, and D. R. Cahoon. 2002. Responses of coastal wetlands to rising sea level. *Ecology* 83: 2869–2877.

Needelman, B. A. and J. E. Hawkes. 2012. Mitigation of greenhouse gases through coastal habitat restoration. In B. A. Needelman, S. Bosak, S. Emmitt-Mattox, and C. Lyons (Eds.) *Creating Resilient Coasts: Coastal Habitat Restoration for Adaptation and Mitigation of Climate Change Impacts*. Restore America's Estuaries, Washington, DC, pp. 49–57.

Nyman, J. A., R. J. Walters, R. D. DeLaune, and W. H. Patrick Jr. 2006. Marsh vertical accretion via vegetative growth. *Estuar. Coast. Shelf Sci.* 69: 370–380.

Oenema, O. 1988. *Early Diagenesis in Recent Fine-Grained Sediments in the Eastern Scheldt*. PhD dissertation, Instituut voor Aardwetenschappen, Rijksuniversiteit Utrecht, Utrecht, the Netherlands, 222pp.

Poffenbarger, H., B. A. Needelman, and J. P. Megonigal. 2011. Salinity influence on methane emissions from tidal marshes. *Wetlands* 31: 831–842.

Rabenhorst, M. C. 1990. Micromorphology of induced iron sulfide formation in a Chesapeake Bay (USA) Tidal Marsh. In L. A. Douglas (Ed.) *Micromorphology: A Basic and Applied Science. Proceedings of the 8th International Working Meeting on Soil Micromorphology*, July 10–15, 1988, San Antonio, TX. Elsevier, Amsterdam, pp. 303–310.

Rabenhorst, M. C. 1997. The chrono-continuum: An approach to modeling pedogenesis in marsh soils along transgressive coastlines. *Soil Sci.* 167: 2–9.

Rabenhorst, M. C. and B. R. James. 1992. Iron sulfidization in tidal marsh soils. In R. W. Fitzpatrick, and H. C. W. Skinner (Eds.) *Biomineralization Processes of Iron and Manganese in Modern and Ancient Environments*. Catena Supplement No. 21, West Germany, pp. 203–217.

Rabenhorst, M. C., J. P. Megonigal, and J. Keller. 2010. Synthetic iron oxides for documenting sulfide in marsh porewater. *Soil Sci. Soc. Am. J.* 74: 1383–1388.

Ranwell, D. S. 1972. *Ecology of Salt Marshes and Sand Dunes*. Chapman and Hall Publishing, London, 258pp.

Redfield, A. C. 1972. Development of a New England salt marsh. *Ecol. Monogr.* 42: 201–237.

Schoeneberger, P. J., D. A. Wysocki, E. C. Benham, and W. D. Broderson. 2012. *Field Book for Describing and Sampling Soils, Version 3.0*. Natural Resources Conservation Service, National Soil Survey Center, Lincoln, NE. ftp://ftp-fc.sc.egov.usda.gov/NSSC/Field_Book/FieldBookVer3.pdf

Soil Survey Staff. 1999. *Soil Taxonomy: A Basic System of Soil Classification for Making and Interpreting Soil Surveys*, 2nd ed. *U.S. Department of Agriculture Handbook No. 436*. U.S. Government Printing Office, Washington, DC.

Soil Survey Staff. 2010. *Keys to Soil Taxonomy*, 11th ed. USDA-NRCS, U.S. Government Printing Office, Washington, DC, 338 pp.

Steers, J. A. 1977. Physiography. In V. J. Chapman (Ed.) *Ecosystems of the World. 1. Wet Coastal Ecosystems*. Elsevier Publishing, Amsterdam, pp. 31–60, 428pp.

Stevenson, J. C., M. S. Kearney, and E. C. Pendleton. 1985. Sedimentation and erosion in a Chesapeake Bay brackish marsh system. *Mar. Geol.* 67: 213–235.

Stevenson, J. C., L. G. Ward, and M. S. Kearney. 1986. Vertical accretion rates in marshes with varying rates of sea-level rise. In D. A. Wolfe (Ed.) *Estuarine Variability*. Academic Press, New York, pp. 241–259.

Stolt, M. H. and M. C. Rabenhorst. 1991. Micromorphology of argillic horizons in an upland/tidal marsh catena. *Soil Sci. Soc. Am. J.* 55: 443–450.

USDA-Natural Resources Conservation Service. 2010. *Field Indicators of Hydric Soils in the United States, Version 7.0*. USDA, NRCS, in cooperation with the National Technical Committee for Hydric Soils, Washington, DC.

Weston, N. B. 2014. Declining sediments and rising seas: An unfortunate convergence for tidal wetlands. *Estuar. Coasts* 37: 1–23.

Widdell, F. 1988. Microbiology and ecology of sulfate- and sulfur-reducing bacteria. In A. J. B. Zehnder (Ed.) *Biology of Anaerobic Microorganisms*. John Wiley & Sons, New York, pp. 469–585.

14

Flatwoods and Associated Landforms of the South Atlantic and Gulf Coastal Lowlands

G. Wade Hurt, Frank C. Watts, and Willie G. Harris Jr.

CONTENTS

Introduction

A landscape is a collection of related natural landforms. It is usually the land surface that the eye can comprehend in a single view. Most landscapes contain many unique and readily identifiable landforms. A landform is a physical, recognizable form or feature on the earth's surface that has a characteristic shape and is produced by natural causes (Tuttle 1975; Soil Survey Staff 1996). This chapter summarizes the soils, hydrology, and related features of flatwoods and other landforms as they occur on the south Atlantic and Gulf Coastal Lowlands landscape (Figure 14.1). Other areas of flatwoods occur in the United States, most notably in southwestern Louisiana and areas parallel to and south and west of the Mississippi and Alabama Blackland Prairies; these will not be discussed.

The term "flatwoods" was coined by Europeans who settled in the southeastern United States to designate the flat areas that support forests of pine (Ober 1954; Abrahamson

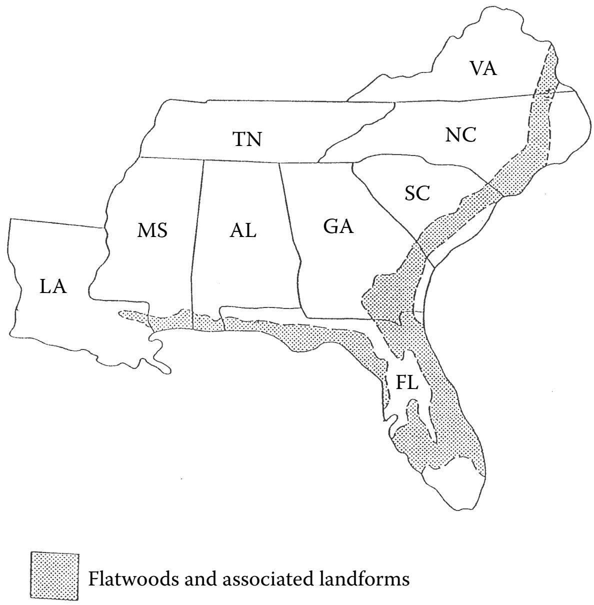


FIGURE 14.1
Major extent of flatwoods and associated landforms of the south Atlantic and Gulf Coastal Lowlands landscape.

and Hartenett 1990). The flatwoods term has long been used to designate a landform and a landscape (Caldwell et al. 1958; Watts et al. 1996). This chapter will refer to flatwoods as landforms as they occur on the south Atlantic and Gulf Coastal Lowland landscapes where they are interspersed with other landforms, such as depressions, flood plains, flats, and rises and knolls (Watts and Carlisle 1997). Flatwoods and associated landforms occur in approximately 50% of Florida (Edmisten 1963; Davis 1967; Abrahamson and Hartenett 1990) and in southeastern Georgia. These landforms comprise relatively smaller portions of the landscapes as they extend northward to Virginia and westward to Louisiana. Subtle differences in local relief and somewhat impervious, geologic strata have primarily influenced the evolution of these different landforms. Flatwoods and associated landforms generally have elevations that are less than 100 m above sea level. The climate is characterized by long, humid, warm summers, and mild winters with annual rainfall of about 1000–1650 mm and average annual temperatures of 13–25°C (Soil Conservation Service 1981).

Soils

The soils of the area classify in seven orders and 11 suborders. Below is a discussion of each order and suborder. This chapter is abbreviated from *Soil Taxonomy: A Basic System of Soil Classification for Making and Interpreting Soils Surveys* (Soil Survey Staff 1999) and *Keys to Soil Taxonomy: Eleventh Edition* (Soil Survey Staff 2010).

Histosols

Histosols are soils that consist of organic materials (more than 12%–18% organic carbon) in at least two-third of the thickness above bedrock and mineral soil layers are less than 10 cm thick *or* are saturated for most of the year and half of the upper 80 cm of soil is organic material. Histosols, as they occur in these landscapes, have organic soil material over mineral soil material or bedrock at varying depths. Sapric is the type of organic soil material that occurs in the area. Sapric soil material (muck texture) is organic material in which most of the plant remains have decomposed such that plant forms cannot be identified (<1/6 fibers after rubbing). Saprists are the Histosols (*ist* is the formative word element from Histosol that appears in the suborder name) that have more sapric material than other, less-decomposed organic (hemic and fibric) materials. These soils are wet most of the year unless artificially drained.

Spodosols

Spodosols are mineral soils that contain a spodic horizon that is 10 cm or more thick. A spodic horizon is a subsurface horizon, usually black to dark reddish brown, in which organic material has accumulated in combination with aluminum and iron due to downward translocation. Depths to spodic horizons vary from <25 cm to 2 m, and thicknesses vary from 10 cm to >1 m; some Spodosols have more than one spodic horizon (Figure 14.2). Bedrock or an argillic horizon (a zone of clay accumulation) may occur at varying depths beneath the spodic horizon (Figure 14.3). Aquods are the Spodosols (*od* is the formative word element from Spodosol that appears in the suborder name) that are wet for extended periods of most years unless they have been artificially drained to reduce the duration of saturation. Orthods are Spodosols that are drier than Aquods.

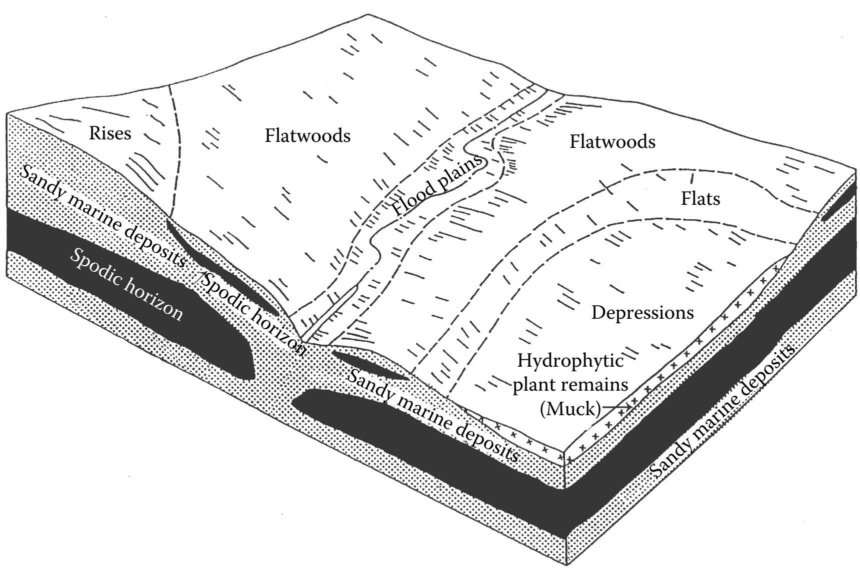


FIGURE 14.2
Common landscape of flatwoods and associated landforms in the northern and western ranges of their occurrence where the flatwoods landform is dominant. This landform pattern is repetitive across the landscape with individual landforms of varying extensiveness. Some landforms may not be present in all landscapes.

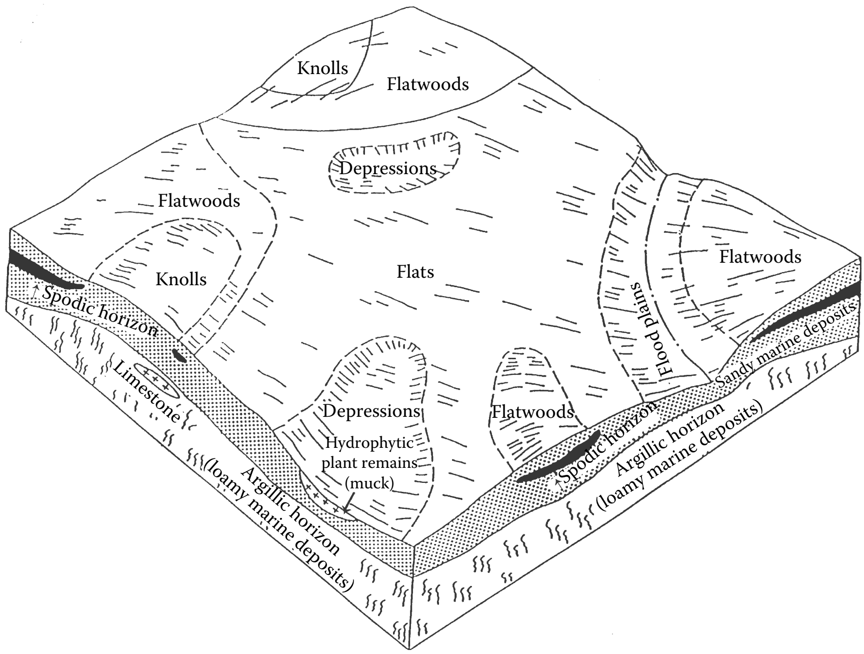


FIGURE 14.3
Common landscape of flatwoods and associated landforms in the southern range of their occurrence where the flats landform is dominant. Similar to the northern and western ranges of occurrence, the landform pattern is repetitive across the landscape with individual landforms of varying extensiveness. Some landforms may not be present in all landscapes.

Ultisols

Ultisols are other mineral soils that have an argillic or kandic horizon and a base saturation of less than 35%. Argillic and kandic horizons are subsurface zones of clay accumulations from the horizon(s) above. Base saturation is the percentage of the soils total cation exchange capacity by base cations (Ca^{2+}, Mg^{2+}, K^+, and Na^+). Depths to argillic or kandic horizons vary from <25 cm to 2 m, and thicknesses vary from <20 cm to >2 m. Bedrock or other material may occur at varying depths beneath the argillic horizon. Leaching associated with the humid climate of the region tends to preferentially deplete base cations unless there is subjacent limestone influence (Ca^{2+} source), in which case soils may classify as Alfisols (see below). Aquults are the Ultisols (*ult* is the formative element) that are wet for extended periods of most years unless they have been artificially drained to reduce the duration of saturation. Udults are the Ultisols that are drier than Aquults; however, the ability of Udults to retain plant-available water varies widely with depth to and thickness of the argillic or kandic horizon.

Mollisols

Mollisols are other mineral soils that have a dark (usually black to very dark gray) mineral surface horizon that is more than 25 cm thick (10 cm is underlain by bedrock) and that has a base saturation of 50% or more. Limestone bedrock and argillic horizons may be present or absent in these soils. Aquolls are the Mollisols (*oll* is the formative element) that are wet for extended periods of most years unless they have been artificially drained to reduce the duration of saturation.

Alfisols

Alfisols are other mineral soils that have an argillic or kandic horizon and a base saturation of 35% or more. Argillic and kandic horizons are subsurface zones of clay accumulations from horizon(s) above. Depths to argillic or kandic horizons vary from <25 cm to 2 m, and thicknesses vary from <20 cm to >2 m. Bedrock or other material may occur at varying depths beneath the argillic horizon. The higher base saturation status of Alfisols may be due to lower leaching intensity or insufficient influence of the underlying limestone to maintain a high percentage of Ca^{2+} on the exchange complex. Aqualfs are the Alfisols (*alf* is the formative element) that are wet for extended periods of most years unless they have been artificially drained to reduce the duration of saturation. Udalfs are the Alfisols that are drier than Aqualfs; however, the ability of Udalfs to retain plant-available water varies widely with depth to and thickness of the argillic or kandic horizon.

Inceptisols

Inceptisols are other mineral soils that have horizon development exemplified by color and/or structure. They have had some horizon development and parent material differentiation but are not enough to class as Spodosols, Ultisols, or other soil orders already described. The limited expression of profile development is most often expressed by differentiating color and structure changes. Inceptisols are most often found on flood plains and marine terraces. Aquepts are the Inceptisols (*ept* is the formative element) that are wet for extended periods of most years unless they have been artificially drained to reduce the duration of saturation.

Entisols

Entisols are all other mineral soils. These soils lack diagnostic horizons in the upper 2 m that might otherwise classify them in one of the soil orders described above. Aquents are the Entisols (*ent* is the formative element) that are wet for extended periods of most years unless they have been artificially drained to reduce the duration of saturation. Psamments are the sandy Entisols. These are the driest soils in the area. They generally have sandy layers to 2 m or more. Both mineral soils and the limnic marl soils (Soil Survey Staff 1999) are classed as Aquents. The marl soils are properly classed in the suborder Aquents and should be classed in the proposed Great Group Limnaquents (Ahrens and Hurt 1999). Limnaquents, as proposed, would include soils composed of mineral materials composed of marl, coprogenous earth, and diatomaceous earth.

Hydric Soils

Hydric soils are defined as soils that formed under conditions of saturation or inundation (flooding or ponding) for periods long enough during the growing season to develop

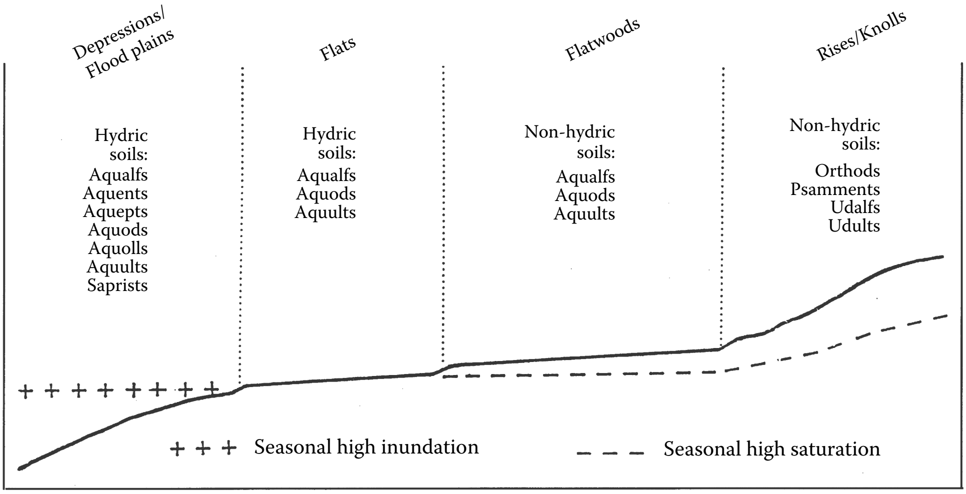


FIGURE 14.4
A common soil toposequence of flatwoods and associated landforms. The depth of seasonal high inundation and depth to seasonal high saturation are shown. Some landforms may not be present in all landscapes.

anaerobic conditions in the upper part of the soil (Federal Register 1994). Saturation is characterized by zero or positive pressure in the soil water with most of the soil pores filled with water. Inundation is characterized by a water table above the soil surface (Soil Survey Staff 2010). Hydric soils have seasonal high saturation and/or inundation for a significant period (more than a few weeks) during the wettest period of the year (Figures 14.4 and 14.5). In the following section on landforms, hydric soils, as they occur on each landform, are identified. These hydric soils have one of the indicators identified in *Field*

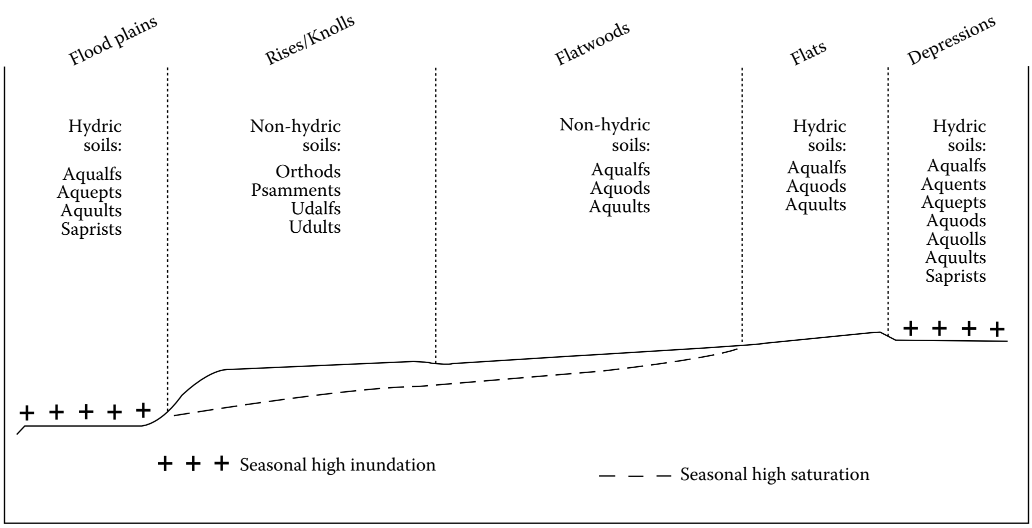


FIGURE 14.5
Most common soil toposequences of flatwoods and associated landforms. The depth of seasonal high inundation and depth to seasonal high saturation are shown. Some landforms may not be present in all landscapes.

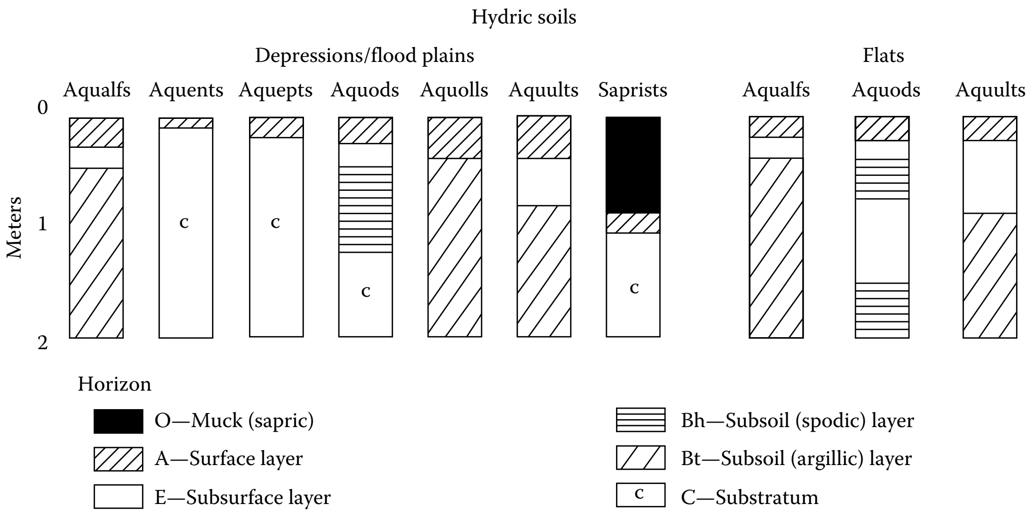


FIGURE 14.6
Idealized pedons that represent the hydric components of the soils associated with the flatwoods and associated landforms of the Atlantic and Gulf Coastal Lowlands landscape. These soils occur on flats and flood plains, and in depressions.

Indicators of Hydric Soils in the United States (USDA NRCS 2010) and *Florida's Hydric Soils: A Guide to Their Recognition (Hurt 2007) in Hydric Soils of Florida Handbook Fourth Edition (Hurt ed.).* Figure 14.6 is a graphic representation of the hydric soils that occur on the wetland landforms. Where the soils are nonhydric, they do not have one of the hydric soil indicators. Figure 14.7 is a graphic representation of the nonhydric soils that occur on the upland landforms.

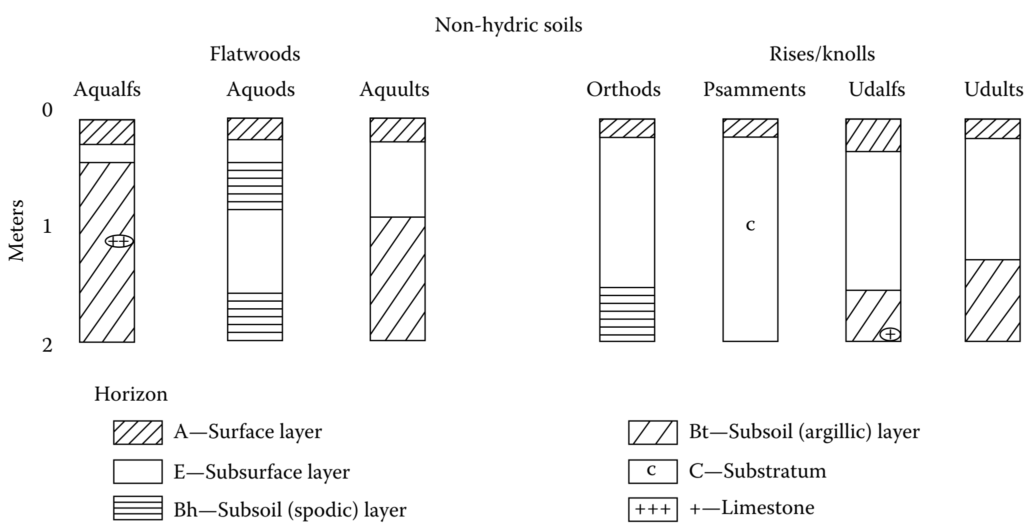


FIGURE 14.7
Idealized pedons that represent the nonhydric components of the soils associated with the flatwoods and associated landforms of the Atlantic and Gulf Coastal Lowlands. These soils occur on flatwoods, rises, and knolls.

Landforms

Flatwoods

Flatwoods landforms (Figures 14.2 and 14.3) are typically broad, nearly level, flat areas with slightly convex and concave relief. The convex relief exists where flatwoods abut flats and depressions. The concave relief exists where flatwoods abut rises and knolls. Flatwoods are 8 cm or more higher than adjacent flats, 30 cm or more higher than adjacent depressions and flood plains, and 15 cm or more lower than adjacent rises and knolls. The only higher closely associated landforms in the coastal lowlands landscapes are rises and knolls and, possibly hillslopes.

Characteristic native vegetation is slash pine (*Pinus elliottii*), longleaf pine (*Pinus palustris*), loblolly pine (*Pinus taeda*), saw palmetto (*Serenoa repens*), gallberry (*Ilex glabra*), wax myrtle (*Myrica cerifera*), chalky bluestem (*Andropogon capillipes*), broomsedge bluestem (*Andropogon virginicus*), and pineland threeawn (*Aristida stricta*). This vegetation normally appears as a shrub-dominated community with an open canopy of pine trees and a sparse herbaceous layer.

The soils that occur on flatwoods characteristically are poorly drained, acid, have a low cation exchange capacity, and a low-to-medium moisture-holding capacity. Flatwoods soils are predominantly not hydric and have seasonal high saturation at a depth of 15–45 cm below the soil surface, although, during periods of high rainfall, they may be episaturated for more than a few days. Aquods, Aquults, and Aqualfs (Figures 14.4 and 14.5) are the most common soils. Aquods are dominant, and often have more than one spodic layer (Figure 14.2) or are underlain by an argillic horizon (Figure 14.3). The nearly level flat to slightly convex relief of flatwoods allows the landform to be readily identified even where it is not vegetated.

During wet seasons, flatwoods transmit water to adjacent, lower-lying flats and depressions through subsurface flow and overland flow and some water to the underlying aquifer through deep seepage. During dry seasons, they receive water from the flats and depressions (Crownover et al. 1995).

Depressions

Depressional landforms (Figures 14.2 and 14.3) are typically sunken parts of the earth's surface; they have concave relief, and do not have natural outlets for surface drainage. They are 30 cm or more lower than adjacent landforms. Depressions commonly occur at the lower elevations of a soil toposequence (Figure 14.4). Most commonly, however, depressions of various sizes are interspersed throughout the landscape (Figure 14.3) as low-lying areas with frequent seasonal high inundation (Figure 14.5).

Characteristic native vegetation is baldcypress (*Taxodium distichum*) and water-tolerant hardwoods. Some depressions are treeless expanses of grasses, sedges, rushes, and other herbaceous plants. Other depressions are dominated by shrubs. Locally, depressions have many names: pocosin in North Carolina, Carolina bay in South Carolina, and freshwater marsh and sawgrass marsh in Florida. Cypress dome and swamp are also common names for depressional landforms throughout the coastal lowland landscapes.

The soils in depressions characteristically are very poorly drained, acid, have a low-to-high cation exchange capacity, a medium-to-high moisture-holding capacity, and are saturated or inundated much of the year. These soils are predominantly hydric. Surface soil horizons are frequently slightly higher in organic matter content as compared with upland soils of similar suborders. Aqualfs, Aquents, Aquepts, Aquods, Aquolls, Aquults, and Saprists are the most common soils. The strongly concave relief and lack of natural outlets allow this landform to be readily identified even where it is not vegetated.

Groundwater flow generally follows the broad elevational gradients of the surface (Crownover et al. 1995). During wet seasons, depressions store water as soil water but mostly as surface water. They transmit some water to the underlying aquifer through deep seepage. During dry seasons, they transmit water back to the surrounding areas of higher-lying flats and flatwoods. This phenomenon is due to the higher evapotranspiration potentials of the flats and flatwoods and is also known to occur in the Carolinas (Lide et al. 1995).

Depressions are important landforms in the landscapes of the south Atlantic and Gulf Coastal Lowlands. They provide a habitat for a large diversity of plants and animals; however, because they frequently occur as small areas dispersed throughout the landscape, many have been drained or partially drained prior to being converted into agriculture and timber production. Where undisturbed, depressions filter pollutants from the surrounding higher-lying landforms.

Flood Plains

Flood plains (Figures 14.2 and 14.3) are constructional landforms built from sediments deposited during overflow and lateral migration of drainageways. Similarly to depressions, flood plains are 30 cm or more lower than adjacent landforms. They also occur at the lower elevations of most soil toposequences (Figures 14.4 and 14.5); however, flood plains, unlike depressions, have natural outlets and have nearly level to concave relief with slightly elevated natural levees adjacent to drainageways.

Characteristic native vegetation is a wide and diverse variety of water-tolerant deciduous hardwoods. A few of the flood plains in these landscapes are treeless expanses of grasses, sedges, rushes, and other herbaceous plants. Some are dominated by baldcypress. Casual observers of flood plains that occur on these coastal lowlands landscapes normally consider them “swamps.”

The soils on flood plains characteristically are poorly to very poorly drained, acid to neutral, have a medium-to-high cation exchange capacity, a medium-to-high moisture-holding capacity, and are saturated or inundated much of the year. Unlike flood plains of many other landscapes, these soils are predominantly hydric. Surface soil horizons are frequently slightly higher in organic matter content as compared to upland soils of similar suborders. Aqualfs, Aquents, Aquepts, Aquods, Aquolls, Aquults, and Saprists are the most common soils. The nearly level, slightly concave relief and presence of natural outlets allows this landform to be readily identified even where it is not vegetated.

During wet seasons, flood plains store water as soil water but mostly as surface water (Figures 14.4 and 14.5). They transmit water to lower elevations on the flood plain, eventually discharging at sea level. Some water is contributed to the underlying aquifer through deep seepage. During dry seasons, such as depressions, flood plains transmit water back to the surrounding areas of higher-lying landforms. This is due to the higher evapotranspiration potentials of the other landforms, and the amount of water is dependent on the elevational gradients.

Flood plains are mostly undisturbed landforms in the landscapes of the south Atlantic and Gulf Coastal Lowlands. They provide a habitat for diverse plants and animals and provide flood protection for the adjacent higher landforms where they are left undisturbed.

Flood plains also filter pollutants from the surrounding higher-lying landforms. Where disturbed, increased flooding of adjacent landforms often results, and pollution reduction is lessened.

Flats

Flats landforms are typically smooth, lack any significant curvature or slope, and evidence little change in elevation, with poorly defined outlets. They are 8–30 cm lower in elevation than adjacent higher flatwoods landforms. Flats commonly comprise relatively insignificant portions of the coastal lowland landscapes (Figure 14.2). They may, however, in the southernmost range of their extent, dominate the landscape (Figure 14.3). Flats have nearly level, slightly concave-to-flat relief. Convex relief exists where flats abut flatwoods and other nonhydric landforms.

Characteristic native vegetation is mixed hardwood and pine with a dense understory of shrubs and saplings in its northern range of occurrence and, in its southern range of occurrence, an open canopy of pine and understory of grasses and/or herbs devoid of shrubs. Flats are commonly known as sloughs in south Florida and as swamps, bayheads, and shrub and pitcher plant bogs in other areas.

Soils on flats characteristically are poorly drained, acid to neutral, have a low-to-medium cation exchange capacity, and a low-to-medium moisture-holding capacity. They are predominantly hydric and have seasonal high saturation at a depth of less than 15 cm below the soil surface (Figures 14.4 and 14.5). During periods of high rainfall, they often have shallow (less than a few cm) inundation for more than a few days. Aquods, Aquults, and Aqualfs are the most common soils. Surface soil horizons are frequently slightly higher in organic carbon content when contrasted to upland soils of similar suborders in the northern range of their occurrence on coastal lowland landscapes. Owing to differential biomass production, the converse is true in the southern range of occurrence.

Flats are the most difficult of the flatwoods and associated landforms for untrained observers to identify because the relief differences between the adjacent higher landforms are subtle. Flats are especially difficult to recognize in their northern range of occurrence because vegetation is usually dense. To the trained observer, the nearly level, concave-to-flat relief and poorly defined outlets are characteristic and observable. Vegetation can also provide a clue for separation in heavily vegetated areas. Flatwoods have the shrub saw palmetto that disappears at the flats landform break to be replaced by other shrubs in its northern range of occupancy and by pineland threeawn grass in its southern range of occupancy.

Groundwater flow generally follows surface elevation gradients. During wet seasons, flats transmit water to adjacent lower-lying depressions via surface and subsurface flow. They transmit water to nonadjacent depressions by lateral flow through the subsurface of the flatwoods soils. During dry seasons, they transfer water back to flatwoods due to the higher evapotranspiration of flatwoods.

Rises and Knolls

Rises and knolls, frequently called ridges, have convex relief. A rise is an imprecise term for a landform that has a broad summit and gently sloping sides, and a knoll is a landform that occurs as a small, low, rounded, and isolated area rising above the lower landforms (Soil Survey Staff 1996). Rises and knolls are typically 15 cm or more higher in elevation than the surrounding wetter landforms.

Characteristic native vegetation is mixed mesic hardwoods and pines forest. This vegetation normally appears as a forest-dominated community with a closed canopy of trees and sparse shrub and herbaceous layers.

The soils that occur on rises characteristically are somewhat poorly drained to moderately well drained, acid, have a low-to-medium cation exchange capacity, and a low-to-medium moisture-holding capacity. Soils on rises and knolls are nonhydric and have seasonal high saturation at depths of greater than 45 cm from the soil surface. Orthods, Psamments, Udualfs, and Udults are the most common soils. The nearly level to gently sloping slightly convex relief of rises and knolls allows the landforms to be readily identified even where they are not vegetated.

Rises and knolls are the most hydrologically isolated of these landforms. They, during wet seasons, contribute water about equally to adjacent lower landforms through lateral flow and to the underlying aquifer through deep seepage. They also contribute some water as overland flow to adjacent lower landforms during high rainfall events. During dry seasons, they neither transmit nor receive from adjacent landforms.

Hillslopes also known as sand hills occur on sloping areas adjacent to flatwoods and extend to either flood plains or lower-lying flatwoods. Hillslopes have similar soils, vegetation, and hydrology as rises and knolls. They are of minor but locally important extent. Owing to their minor extent, they are not represented in Figure 14.4.

Summary

Soils of the flatwoods and associated landforms of the south Atlantic and Gulf Coastal Lowlands landscape classify into the following orders: Histosols, Spodosols, Ultisols, Mollisols, Alfisols, Inceptisols, and Entisols. Also recognized are 11 suborders: Aqualfs, Aquents, Aquepts, Aquods, Aquolls, Aquults, Orthods, Psamments, Saprists, Udalfs, and Ulults. Soils of the flatwoods are dominantly Aquods. All other associated landforms lack a dominant soil suborder.

The south Atlantic and Gulf Coastal Lowland landscapes commonly have the following landforms: flatwoods, depressions, flood plains, flats, and rises, knolls, and (rarely) hillslopes. Depressions, flood plains, and flats most commonly have hydric soils. Depressions and flood plains primarily function as discharge wetlands, and flats function as flow through/discharge areas to other wetlands. Flatwoods, rises, knolls, and hillslopes characteristically have nonhydric soils. These landforms function as recharge areas to wetlands and to underlying aquifers.

Each of the flatwoods and associated landforms of the Atlantic and Gulf Coastal Lowland landscapes has a characteristic shape and relief that make them readily identifiable with or without the presence of vegetation.

References

Abrahamson, W. G. and D. C. Hartenett. 1990. Pine flatwoods and dry prairies. In R. L. Myers, and J. J. Ewel (Eds.) *Ecosystems of Florida*. University of Central Florida Press, Orlando, FL, pp. 103–149.

Ahrens, R. J. and G. W. Hurt. 1999. Proposed Limnaquents Great Group, USDA, NRCS, Lincoln, NE. *Soil Surv. Horiz.* 40(1): 13–16.

Caldwell, R. E., O. C. Olson, J. B. Cromrite, and R. G. Leighty. 1958. *Soil Survey of Manatee County, Florida.* USDA, Soil Conservation Service, U.S. Government Printing Office, Washington, DC.

Crownover, S. H., N. B. Comerford, D. G. Neary, and J. Montgomery. 1995. Horizontal groundwater flow patterns through a cypress swamp—Pine flatwoods landscape. *Soil Sci. Soc. Am. J.* 59: 1199–1206.

Davis, J. H. Jr. 1967. *General Map of the Natural Vegetation of Florida*. Institute of Food Agricultural Experiment Station, Circ. S-178, University of Florida, Gainesville.

Edmisten, J. E. 1963. *The Ecology of the Florida Pine Flatwoods.* PhD thesis, University of Florida, Gainesville.

Federal Register. 1994. *Changes in Hydric Soils of the United States*. National Archives and Records Administration and U.S. Government Printing Office, Washington, DC, July 13.

Hurt, G. W. 2007. *Florida's Hydric Soils: A Guide to Their Recognition.* In G. W. Hurt (Ed.) *Hydric Soils of Florida. Fourth Edition.* Florida Association of Environmental Soil Scientists, Gainesville, FL.

Lide, R. F., V. G. Meentemeyer, J. E. Pinder III, and L. M. Beatty. 1995. Hydrology of a Carolina bay located on the upper coastal plain of western South Carolina. *Wetlands* 15: 47–51.

Ober, L. D. 1954. *Plant Communities of the Flatwood Forests in Austin Cary Memorial Forest.* MS thesis, University of Florida, Gainesville.

Soil Conservation Service. 1981. *Land Resource Regions and Major Land Resource Areas of the United States.* USDA, SCS, Agriculture 296, U.S. Government Printing Office, Washington, DC.

Soil Survey Staff. 1996. *National Soil Survey Handbook.* USDA, NRCS, U.S. Government Printing Office, Washington, DC.

Soil Survey Staff. 1999. *Soil Taxonomy: A Basic System of Soil Classification for Making and Interpreting Soil Surveys,* 2nd ed. *USDA Agricultural Handbook 436*. U.S. Government Printing Office, Washington, DC.

Soil Survey Staff. 2010. *Keys to Soil Taxonomy,* 11th ed. USDA, NRCS, U.S. Government Printing Office, Washington, DC.

Tuttle, S. D. 1975. *Landforms and Landscapes*. Wm. C. Brown Company, Dubuque, IA.

USDA NRCS. 2010. Field indicators of hydric soils in the United States, Version 7.0. In Vasilas, L.M, G.W. Hurt, and C.V. Noble (Eds.) *USDA, NRCS, in Cooperation with the National Technical Committee for Hydric Soils*.

Watts, F. C., T. S. Bowerman, R. A. Casteel, D. Hinz, A. Jenkins, J. C. Remley, T. J. Solem. et al. 1996. *Soil Survey of Baker County, Florida.* USDA, NRCS, U.S. Government Printing Office. Washington, DC.

Watts, F. C. and V. W. Carlisle. 1997. Soils of flatwoods and associated landforms of Florida. Soil and crop sciences of Florida. *Proceedings* 57: 59–65.

15

Saline and Wet Soils of Wetlands in Dry Climates

Aaron J. Miller, Janis L. Boettinger, and Jimmie L. Richardson

CONTENTS

Introduction

Wet soils occur in arid and semiarid climates typical of central and western North America. Though not widespread, wet soils in dry and seasonally dry climates perform important ecological and environmental services in these regions providing wildlife habitat and corridors and helping to scrub surface and groundwater of natural and anthropogenic pollutants. Though their function and significance may be similar to humid wetland counterparts, their morphology and characteristics can be considerably different. It is therefore important to understand the factors and processes involved in their distribution and formation.

Arid wet soils and wetlands are commonly found at groundwater discharge sites such as flood plain edges, pluvial lake playas, prairie playas, artesian spring areas, and areas associated with irrigation practices. Some of these landforms, such as flood plains, only experience a seasonal high water table for a month or less during an average year. Other areas, such as playas, may see a return interval for wet soils only a few times per decade. Many of the wet soils in dry climates have formed under high evapotranspiration rates. When coupled with low effective precipitation and poor drainage, this environment leads to concentrated salt accumulations. Therefore, the hydrogeomorphology, geological sources of the water and sediment, and evapo-concentration of salts are important factors to consider when working with wet soils of arid climates.

Wet soils in arid regions commonly contain carbonates, gypsum, and sometimes more labile salts. Biomass accumulation is limited in many saline wetlands, and soil organic matter, not surprisingly, can be low. The accumulation of soluble salts and low inputs of organic matter probably hinder biogeochemical processes expected in wet soils, such as chemical reduction of Fe^{3+}. This is due to the chemical relationship between alkalinity and redox potential where an increase in the pH requires a lower Eh to be reached to achieve similar anaerobic conditions (Chapter 4). These processes impede the formation of redoximorphic features in saline wet soils. However, productivity of some saline wetlands such as salt marshes is high, and detritus from some of these landscapes feed abundant wildlife. Bai et al. (2010) partially attributed higher nitrification rates of marsh soils in saline wetlands to higher pH values of closed wetlands.

Boettinger (1997) summarized the geographic distribution, parent material, landform, and vegetation of saline and wet soils mapped as Aquisalids (formerly Salorthids) in soil surveys of the United States. Early USDA soil surveys sometimes mapped wet areas in dry climates as miscellaneous areas due to the small amount of vegetative cover (<10%), seasonal presence of standing water, thinness of soil, or short residence time of sediments or other soil materials (e.g., even freshwater wetlands were placed in miscellaneous areas through the 1970s). As the importance of wetland resources is brought to the forefront of conservation priorities, more hydric soil components are becoming established and more soil detail will be mapped. For instance, a movement to update and remap the playa wetlands of northwest Texas and eastern New Mexico has been underway for some time beginning in 2010. We can expect to see a much improved dataset for hydric soils in these areas in the near future.

This chapter summarizes the factors and processes involved in the distribution and formation of saline and wet soils and wetlands of dry climates. These include environmental setting, pedogenesis, and the influence of human activity. The challenges of separating wet, saline soils into hydric and nonhydric status are also addressed.

Environmental Setting

Hydrology

Wet soils of dry climates occur in areas of specific water discharge or accumulation. Open-flow systems (areas through which streams flow), such as flood plains, have points of discharge where water can accumulate (see Chapter 3). Closed-basin systems that lack external drainage, such as playas, serve as sinks for surface runoff and groundwater discharge. Wet soils can occur in the vicinity of springs in either open- or closed-flow systems. In addition, wet soils can be created in areas irrigated for agriculture, areas around drainage ditches where water accumulates, or areas where water seeps from drainage ditches or irrigation delivery systems.

Soils with internal drainage, such as playas, serve as both zones of discharge and zones of recharge (Figure 15.1). These soils can be saline, hydric, or potentially both. These playas are occasionally barren but more commonly have vegetative cover. As a low point on the landscape, the surrounding area concentrates water into the playa through runoff and through flow; however, playas only lose water through evaporation, transpiration, and recharge. In the case of recharge, surface water can become groundwater that helps replenish deeper aquifers. This happens at an accelerated rate after a long dry spell when the playa clays may be deeply cracked. The cracks act as direct pipelines to channel water deeper into the ground (Figure 15.1).

Climate

Arid and semiarid regions typically receive a limited amount of precipitation distributed throughout the year, and can be seasonally droughty. In addition, these climates are

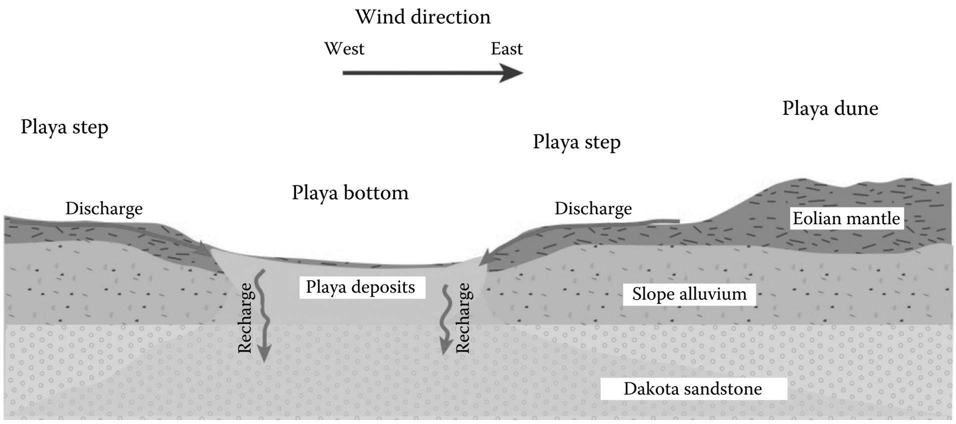


FIGURE 15.1
Diagram of a typical prairie playa in eastern New Mexico. A portion of the annual soil moisture will infiltrate the relatively flat surfaces of the plateau or plains and can reemerge as discharge into depression areas such as playas. The depressions then serve as recharge for deeper water tables.

typified by high evapotranspiration rates due to high amounts of incoming solar radiation, and low humidity. Because annual evaporation exceeds effective precipitation, the soils in these regions frequently accumulate calcium carbonate ($CaCO_3$), gypsum, and salts more soluble than gypsum. Occasionally, saline and wet soils occur in subhumid areas with a pronounced dry season or on landscape positions that have a source of salts and restricted drainage.

The seasonal timing of precipitation plays a strong role in how the ecosystem utilizes moisture. During warm summer months when biological activity could thrive may not coincide with the period of available water. In Mediterranean-type climates with xeric soil moisture regimes, total precipitation may be plentiful enough to wet the soil profile but the drought season occurs when water demand is highest and the soils can dry to greater depths. In regions with ustic soil moisture regimes such as the southwestern states with a monsoonal summer rainfall pattern, precipitation occurs during the growing season. Therefore, the soil may not dry as deep or severe because there is moisture replenishment during the period of the highest evapotranspirative demand. This moisture-timing effect will help determine the severity of salt accumulation, especially during the growing season when plants are most vulnerable and soil biogeochemical reactions are throttled up. Furthermore, in summer months, when heat powers the capillary pumps that pull salts dissolved in soil moisture to the surface, summer rains can help flush these salts and lower the electrical conductivity (EC) in the A horizons. In some soils, this process may be enough to prevent the formation of a salt crust that would otherwise inhibit the growth of most plants.

Geomorphology and Salt Source

Playas and lake basins represent the most common geomorphic setting for saline and wet soils and wetlands of dry climates and have been studied extensively. For example, pluvial lake playa systems are common throughout the Basin and Range physiographic province of western North America and provide the setting for almost half the soils mapped as Aquisalids in the United States (Boettinger 1997). Playas are also found in the shortgrass

FIGURE 15.2
(**See color insert.**) Examples of playas near Las Vegas, New Mexico; one dry (left) and one filled with water (right). The playa now filled with water might experience a period of unvegetated soil following the dryout period as much of the perennial grass and other species will have undoubtedly been drowned. The playa that remains mostly unponded will likely remain vegetated and will provide a productive pasture via good run-on and storage of water. (Photo provided by A.J. Miller, NRCS, NM.)

prairies of the western Great Plains Province (Figure 15.2). These are formed through a combination of wind, wave, and dissolution processes and fill with water through precipitation and runoff (Bolen et al. 1989). Playas and lake plains in arid and semiarid regions range in salinity from nearly salt-free intermittent ponds to hypersaline salt flats. These landforms occur in basins that lack outlets or do not have integrated drainage systems. Because of the closed nature of the basins and impact of long-term climatic variation, the amount of water in these basins has varied widely over recent geologic time. During pluvial cycles, lakes of significant depth and areal extent exist in these basins. As the climate becomes warmer and drier, a greater water deficit will cause lake levels to drop or disappear entirely.

Initial salt concentration in closed-basin pluvial lake water was probably low. Dissolved products in these lakes were primarily derived from chemical weathering of minerals in the hydrologic and geologic source areas. In the Great Basin, geologic sources are dominantly sedimentary rocks in the east, grading to igneous and metamorphic rocks in the west. Other sources include rain, snow, and aerosol particulates. As lake water evaporated, dissolved products were left behind to concentrate in sediments. The characteristics of playa sediments are determined by the balance between clastic sedimentation during pluvial stages and salt deposition between pluvial stages. Soils formed during dry, nonpluvial periods are subject to diurnal and seasonal climatic fluctuations, which can cause fluctuations in the presence and mineralogic composition of salts (e.g., Keller et al. 1986; Eghbal et al. 1989).

On a smaller scale, saline and wet soils occur in recharge–throughflow–discharge wetland systems, typical of a glaciated terrain in subhumid climates of the northern Great Plains of central North America (Arndt and Richardson 1989; Chapter 3). These landscapes are characterized by little vertical relief (<20 m) on expansive plains of till derived from dolomitic and sulfur-rich marine sedimentary rocks that are underlain by bedrock, which restricts downward movement of water. Arndt and Richardson (1989) found that water is subject to increasing evaporation and concentration as it moves from recharge areas to intermediate throughflow areas, and finally to the local sink in the discharge area. In general, soils in recharge wetlands are nonsaline, whereas soils in discharge wetlands are often saline. In the northern prairie region of the Great Plains, water in closed-basin wetlands interacts with subsurface pedogenic gypsum to maintain a moderate salinity in these systems. The gypsum has been added to these systems since deglaciation of the region as sulfate is transferred from the surface water to underlying sediments during the drying phase of the wetland. However, only part of this sulfate recycles back to the surface water during wetting periods (Heagle et al. 2013). This hydrologic relationship has also been observed in the southwestern Great Plains of New Mexico and Texas forming over sandstone plateaus and some Ogallala sediments where glaciation has not occurred.

Saline wet soils and wetlands also occur near artesian springs in either open- or closed-flow systems. In closed-flow systems, water and salt from artesian springs may not be easily differentiated from overland sources. In contrast, the contribution of saline groundwater from artesian springs is more easily distinguished in open-flow systems. For example, the saline water source for about 600 ha of the Cache series (fine, mixed, mesic, and semiactive Typic Aquisalid) in Cache Valley, northern Utah, is artesian springs. These springs occur in a north–south trend, apparently at the southern termination of the Dayton fault (S. U. Janecke, personal communication, 1998). The EC of the spring water is about 5 dS m^{-1} (U.S. Geological Survey 1970). Although the very slowly permeable soils in this saline wetland area are externally drained, salts have accumulated in these soils as evidenced by EC (saturated paste) values of up to 47 dS m^{-1} (Erickson and Mortensen 1974).

Saline wetlands can also occur on coastal flats or terraces that are periodically inundated by salt water. Seven of the 24 soil series mapped as Aquisalids in the United States occur in these environmental settings (Boettinger 1997). Tidal water and storm surges can quickly provide both water and dissolved salts. If these areas are in warm and/or seasonally dry climates with high evaporation rates, saline wet soils may form.

Some saline and wet soils have been saturated and salinized due to human activity. Seepage from drainage ditches along roads has caused the water table to rise in adjacent areas, and salts are concentrated by evaporation (Skarie et al. 1987). Irrigated agriculture in arid and semiarid regions is probably the major anthropogenic cause of saline and wet soils and artificial wetlands (Boettinger 1997). Seepage from irrigation water canals, such as those constructed in marine shale in central Utah, is related to salinization of adjacent soils. Irrigation may cause ponding in depressional areas and raise the local water tables. These areas of water accumulation may salinize as water is evaporated and salts are concentrated.

Pedogenesis

Saline and wet soils in dry climates, such as Aquisalids, may form if the following three conditions are met: (1) there must be a source of water; (2) there must be a source of ions or dissolvable minerals that can be translocated by water; and (3) there must be a process of solution concentration (i.e., evapotranspiration exceeds precipitation).

A key pedogenic process involved in the formation of saline and wet soils in dry climates is salinization, or the accumulation of evaporite minerals. Hardie and Eugster (1970) clearly explained evaporite accumulation and the various mineral assemblages that result (Figure 15.3). Their paper is based on the sequences of mineral accumulation that occur by progressive salt precipitation via evaporation in closed basins. Whittig and Janitzky (1963), Gile and Grossman (1979), Gumuzzio et al. (1982), Last (1984), Keller et al. (1986), Arndt and

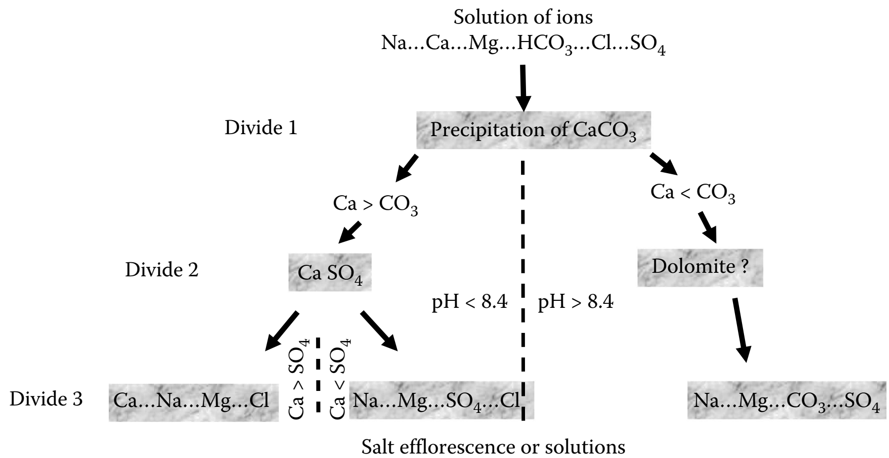


FIGURE 15.3
An illustration of a closed-basin brine evolution model for soils. Starting with a series of common ions found in soil solutions, calcite precipitates followed by gypsum in one pathway. The precipitation of equal molar amounts of ions tends to concentrate the ion with the largest solution concentration and all the unaffected ions in solution. (Modified from Hardie, L. A. and H. P. Eugster. 1970. *Miner. Soc. Am. Spec. Paper* 3: 273–290.)

Richardson (1989), and Eghbal et al. (1989) related evaporite mineral formation and salt accumulation in general to pedogenesis.

The Hardie–Eugster (1970) evaporite sequence is based on "chemical divides," and depends on the initial ionic composition of the solution undergoing evaporation. Salts are ionically bonded, binary compounds that precipitate from the solution when the product of the activity of the ions in solution equals or exceeds the solubility product for that particular mineral. A dilute solution becomes more concentrated with ions as water evaporates. The first salt to precipitate from the solution is calcite, a mineral of the compound $CaCO_3$, which has one mole of Ca^{2+} that combines with one mole of CO_3^{2-} to produce one mole of $CaCO_3$. Calcite precipitation creates a "chemical divide." In solution, it is highly unlikely that Ca^{2+} and CO_3^{2-} will occur in exactly equal portions. If Ca^{2+} is dominant, then, the available CO_3^{2-} in the concentrating solution is quickly consumed during precipitation of calcite. The resulting solution will be enriched with Ca^{2+} with respect to carbonate. If CO_3^{2-} is dominant, the Ca^{2+} is consumed as calcite is precipitated, causing the resulting solution to be enriched with CO_3^{2-} with respect to calcium. We believe there are two distinct systems of chemical divides or pathways of pedogenesis, "alkaline" and "gypsiferous."

If sulfate (SO_4^{2-}) is present in the solution enriched with Ca^{2+}, the excess Ca^{2+} combines next with SO_4^{2-} and gypsum ($CaSO_4 \cdot 2H_2O$) precipitates. This direction is termed the "gypsiferous" path of pedogenesis. Precipitation of gypsum creates another chemical divide, allowing either Ca^{2+} or SO_4^{2-} to enrich the solution, depending on whether Ca^{2+} or SO_4^{2-} was in higher molar concentration before gypsum formation (Figure 15.3).

The opposite side of the first divide has the concentration of CO_3^{2-} being greater than the concentration of Ca^{2+}. As CO_3^{2-} concentration increases, the pH increases. This direction is termed the "alkaline" path of pedogenesis. The ratio of bicarbonate to carbonate is 1:1 at a pH of 10.33 (Lindsay 1979). In the Sacramento Valley of California, it was shown that $NaHCO_3$ was produced during periods of high water table due to reduction in the center of a salty basin. As these HCO_3^- laden waters migrated toward the basin rim, evapotranspiration resulted in a decrease in dissolved CO_2 causing $NaCO_3$ to be formed and the pH to increase (Whittig and Janitzky 1963). Heat, agitation, or other action can also deplete the carbon dioxide and bicarbonate from the solution. When bicarbonate is depleted, the carbonate increases, as does the pH. Note that carbonate is nearly nonexistent until calcite starts to precipitate but increases logarithmically with a decrease in bicarbonate. A 10× increase in carbonate increases the pH by one unit. During the winter, pH was <6.0, whereas in summer, pH was >9.0. Warm weather and wave agitation caused carbon dioxide to degas from water and increase the carbonate at the expense of bicarbonate, clouding the water with fine precipitates of calcite.

Degassing of carbon dioxide during warm weather, high evaporation, and perhaps some wave action enhances calcite precipitation and increases alkalinity. We expect two distinct pathways of pedogenesis based on climate: (1) sodium carbonate-enriched, highly alkaline systems (pH > 8.5); and (2) $CaCO_3$ systems with moderately alkaline conditions where pH values range from 7.8 to 8.7 (Arndt and Richardson 1989; Steinwand and Richardson 1989). The geochemistry of these chemical interactions is discussed in some detail in Arndt and Richardson (1992). Alkaline Natraquolls associated with ponds should be expected in warmer climates where evaporation exceeds precipitation. Calciaquolls should occur in cooler climates. The alkaline pathway consumes its Ca in calcite precipitation leaving no Ca remaining to form gypsum. Complete calcite precipitation depletes Ca but has no impact on Mg. Dolomite ($CaMg[CO_3]_2$) should be the next evaporite to form along the alkaline branch of the Hardie–Eugster (1970) chemical divides system.

Last and DeDeckker (1992) measured several meters of evaporites in Lake Beeac in Australia and found that the sediments were largely authigenic dolomite and magnesite ($MgCO_3$). Sherman et al. (1962) observed dolomitic Bk horizons in soils of glacial Lake Agassiz. Kohut et al. (1995) observed authigenic dolomite in Alberta soils. After dolomite forms, either Mg or carbonate will be in short supply, creating another chemical divide. The low carbonate pathway contains Mg chloride or sulfate, whereas the high carbonate (low Mg) pathway contains Na carbonates, chlorides, and sulfates.

Further evaporation, whether it occurs along the alkaline or gypsiferous pathway, results in the precipitation of highly soluble evaporite minerals. The presence of these labile minerals may fluctuate diurnally and seasonally, depending on fluctuations in environmental conditions such as temperature, relative humidity, amount of solar radiation, and amount and intensity of precipitation events. The most important point here is that with evaporation, a sequence of evaporites forms and each of the preceding evaporites affects the next generation of evaporites by removal of solute supply. Initial solution composition, evaporation, and precipitation effectively control the species of evaporates formed (Hardie and Eugster 1970; Last 1984; Keller et al. 1986; Arndt and Richardson 1992).

Soil Morphology

Aquisalids and other saline and wet soils usually have A horizons that are enriched in organic matter with respect to the underlying strata. Under sparse vegetation, these horizons may be difficult to locate or may not exist at all. Boettinger (1997) stated that soils that lack vegetation or are very sparsely vegetated, such as soil in the intermound areas of mound–intermound complexes in saline lake basins in the Great Basin in Nevada and

FIGURE 15.4
(**See color insert.**) Salt crust formed by evaporation of water from a floodplain in Salt Creek of the Tularosa Basin in southern New Mexico. The gypsum salt crust is commonly coated on the underside by halophytic algae. (Photo by G. Cates, NRCS, NM.)

western Utah, may lack A horizons. Thus, C horizons are probably present at the soil surface. Soluble salt crystals and salt crusts on the soil surface (Figure 15.4), or Az or Cz horizons can disappear quickly during rain events but can reappear during times of high evaporation.

Conventions for describing Bz horizons are in flux at the time of this chapter, but generally require the presence of salts more soluble than gypsum. The presence or absence of soil structure is applied inconsistently when naming subsurface horizons.

Epiaquerts, such as the Randall series indicative of playa soils in western Texas, or Haplusterts such as the Stanley series of east-central New Mexico are common soil taxa for playas (USDA-NRCS Official Soil Series Descriptions). The Randall series is frequently flooded whereas Stanley series is rarely flooded. Both soils are high in clay and have characteristics of Vertisols such as slickensides and wedge-shaped aggregates. However, the Randall soil has little-to-no secondary salts described in the profile whereas the Stanley soil has visible salt crystals starting at 8 cm (Table 15.1). The Randall soils also meet hydric indicators A11 and F6; the Stanley soils do not meet an indicator and are presumed to be not hydric.

Additionally, lack of vegetation for an area should not be used as a criterion that an earth substrate is not soil. Both the potential for plant growth and the presence of bacteria or other microbial flora could be present as well as other signs of life that are commonly overlooked. If the substrate has a structure or other pedogenic features, or has formed by processes that can be considered pedogenic, it should be described as soil.

As a general rule, classic redoximorphic features are likely to be poorly expressed in saline and wet soils. Boettinger (1997) reviewed the morphological properties and depths to redox concentrations present in the typical pedons of Aquisalids. Only 10 of the 24 series of Aquisalids had redoximorphic features in the upper 30 cm. Boettinger (1997) listed several possible reasons, but we speculate that the predominate reason is the lack of carbon as an electron donor and microbial energy source that prevents the Eh from decreasing to a level where iron or manganese can reduce. Often, secondary salts can

TABLE 15.1

Chemical Properties of Playa Soils in Two Mapping Units

County	Depth (in.)	Clay (%)	Soil pH	Calcium Carbonate (%)	Gypsum (%)	Salinity (mmhos cm^{-1})	Sodium Absorption Ratio
Santa Fe Co., NM	*Kwahe–Stanley complex, 0%–1% slopes, ponded (Stanley data shown)*						
	0–3	60–80	7.9–8.5	15–25	0	0–2	0–4
	3–17	"	"	"	0	"	"
	17–28	"	"	"	0	2–4	"
	28–38	"	"	"	0	"	"
	38–56	"	"	"	0–1	4–8	"
	56–101	"	"	"	1–2	"	"
Deaf Smith Co., TX	*Randall clay, 0%–1% slopes, frequently ponded*						
	0–9	55–70	6.1–7.8	0	0	0–1.5	0–1
	9–17	"	6.6–7.8	"	0	0–1	"
	17–62	"	6.6–7.8	"	0	"	"
	62–80	"	6.6–8.4	0–15	0	"	"

Source: Data from the National Soil Information System, USDA-NRCS.

mask the colors from redox chemistry, making a positive hydric indicator tricky to identify (Figure 15.5). It is therefore mandatory that redox concentrations are also identified in the profile.

A soil color change upon oxidation, where carbonates, gypsum, and/or soluble salts are present, may be another good indicator that redox chemistry is actively present. Under these conditions, hydric soil indicators may not be apparent or could be masked by the evaporite minerals. In a wet soil in the Tularosa Basin of New Mexico, soluble manganese mixes with dissolved sulfate in areas with a plentiful supply of gypsum to form MnS (Figure 15.6). These mutable colors can be detected by a dark mineral coating or layer that disappears immediately upon contact with 3% H_2O_2. According to the National Technical Committee for Hydric Soils, which oversees the hydric soil indicators, new indicators such as those discussed above can be proposed along with a clearly formed set of identifiable diagnostics that would be submitted for testing, along with supporting data, before its adoption (USDA NRCS 2010).

Many soils rich in carbonates, gypsum, and/or soluble salts will change color upon exposure to air. Yellowing of the hue, such as 2.5Y changing to 5Y, or a change in chroma, such as from 3 to 2, upon drying may be due to altered hydration or further precipitation of evaporites (Boettinger 1997). Upon drying, changes in the size and structure of crystals can also change the appearance and perhaps color patterns of soils (Last 1984; Keller et al. 1986). These color changes can be confused with those of a reduced matrix where reduced iron (Fe^{2+}) in the solution quickly precipitates (Fe^{3+}) upon exposure to air.

Some saline wet soils can develop very unusual color patterns. Timpson et al. (1986) noted natrojarosite concentrations with bright yellow colors and red gypsum crystals associated with sulfatic soils in saline seeps in North Dakota. In these soils, pH can be low because of sulfide oxidation. The peculiar color association that occurs in sodic–saline seeps with jarosite should be considered as exceptional but not rare because Cretaceous shales and Tertiary lignites are very widespread.

FIGURE 15.5
(**See color insert.**) A Typic Aquisalid in the Carrizo Plain, California. In this playa landform position, small amounts of organic matter can accumulate in the playa surface, enough to feed anaerobic soil microbes. Notice the low chromas of the surface horizons (between ribboned pins) indicative of redoximorphic conditions. Also visible in the first horizon is finely disseminated secondary salts that may mask the colors needed to identify hydric indicators. This soil does not have enough organic matter at the surface to form a distinct A horizon; directly below the salt crust is a Cz1 horizon. (Photo by A. J. Miller NRCS, NM.)

Conclusions: Implications of Arid Climates on Hydric Soil Indicators

When attempting to identify hydric soils in dry climates, hydric soil indicators (USDA NRCS 2010) are often difficult to use. Hydric soil indicators, for the most part, have not been developed with much consideration of soil morphology related to Aridisols. The challenges to development of hydric soil morphology in arid climates primarily arise due to the presence of salts; the low amounts of organic matter; contemporary eolian deposits or alluvial sedimentation; and disturbance of the soil surface due to agriculture or urban activities.

Most floodplains along major aridland streams are intensively utilized for irrigated and nonirrigated crop cultivation. These practices have lowered the soil organic matter content to the point where it can be a limiting factor in anaerobiosis, an important process in the creation of hydric soil morphology (Chapter 7). This, coupled with historic plowing of the surface soil horizons, blurs the evidence of hydric soil morphology or else makes its identification very difficult. Further complications to identifying indicators can occur when floodplain sediments are sourced from high-chroma parent materials. Such minerals have strong red pigments from iron oxides locked deeply within the mineral lattice or structure, such as red chert, and are resistant to the formation of low-chroma colors.

Playa soils in areas where agriculture dominates the landscape are commonly exposed to high rates of sedimentation. Attempts to reduce the rates of sedimentation by installing

FIGURE 15.6
(**See color insert.**) Gypsic Aquisalid observed in the Tularosa Basin of New Mexico. In a layer just below the surface, darkly colored Mn^{2+} accumulations, likely as manganese monosulfide (MnS), form as mineral coatings around ped surfaces, an indicator that both dissolved manganese and a source of sulfur exist. Precipitates of this reduced form of Mn can be easily verified in the field by applying 3% H_2O_2 and watching the minerals dissolve rapidly, leaving no trace of the black MnS. (Photo by A. Miller, NRCS, NM.)

conservation plantings and buffers have reduced the frequency of flooding in the playas themselves (Cariveau et al. 2011). Where sedimentation is measurable, any layers with redox morphology are likely to be buried below the diagnostic depths of hydric indicators. Furthermore, in areas where practices to reduce sedimentation are causing lower flood return cycles, the soils may no longer experience saturated conditions long enough to cause reduction to occur. Playa soils that have vertic properties, such as the Stanley series (Aridic Haplusterts), form deep cracks during the dry seasons and fill with sediment from eolian or overland flow events. These fresh materials become mixed into the upper part of the soil and dilute the hydric morphology (Figure 15.7).

With no data published on the saturation and redox potential of wet, saline soils, Boettinger (1997) concluded that some highly saline soils might not experience reducing conditions. The major limitations to reducing conditions in soils are the chemical or physical constraints on microbial activity. Very negative osmotic potential, low organic C and N supplies in barren or sparsely vegetated soils, as well as salinity, alkalinity, and induced nutrient deficiencies were all cited as potential limitations to microbially mediated reduction during periods of saturation.

However, data show that saline and wet soils subject to saturation within 30 cm of the surface for several weeks during the growing season can experience reducing conditions. Sutcliffe (1999) monitored water table depth, pH, EC, temperature, and dissolved oxygen (DO) in four soils in a hillslope catena formed in marine shale of the Mancos Formation in central Utah. These soils, ranging from nonsaline and never saturated to various degrees of

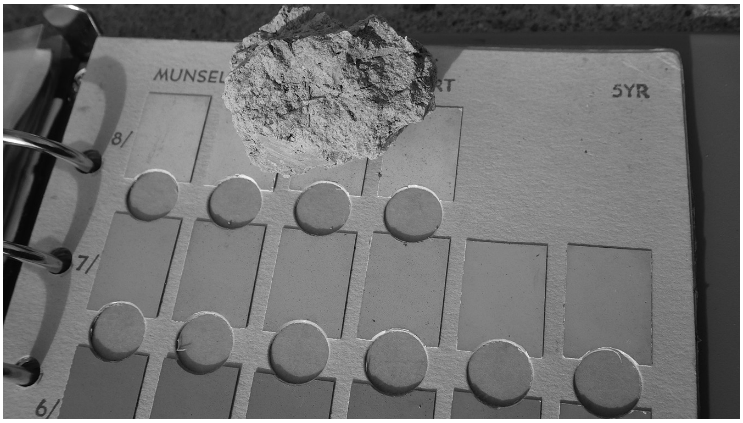


FIGURE 15.7
(**See color insert.**) This ped sample from the Stanley series shows a combination of two materials due to vertic mixing. The light greenish color (5YR 8/1) is from reduction in the soil profile, and the redder hues (5YR 7/2) are from material that is washed or blown into the profile when the soil is dry and deeply cracked. (Photo provided by A. Miller, NRCS, NM.)

salinity (slight to strong) and saturation, were affected by seepage from upslope irrigation canals. One soil in the catena, a Typic Halaquept with an EC of about 20 dS m^{-1}, became anaerobic as the soil warmed and remained anaerobic until the water table dropped to about 30 cm (Table 15.2). This soil develops a fluffy salt crust for part of the year, similar to that shown in Figure 15.4 but does not express any redoximorphic features in the upper part of the soil.

TABLE 15.2

Physical, Chemical, and Redox Properties of a Typic Halaquept Affected by Seepage of Irrigation Water through Mancos Shale

Date	Depth to Water (cm)	pH	EC (dS m^{-1})	Temperature (°C)	DO (mg L^{-1})	Eh (mV)	Anaerobic Eh Threshold	Technical Standard Limit
5/19/98	9	8.1	21.2	12.6	4.0	261	231	109
5/19/98	9	8.1	21.2	12.6	4.0	261	231	109
6/3/98	15	7.8	20.4	14.3	2.2	37	250	127
6/17/98	14	7.7	20.5	13.1	1.9	–23	254	133
7/2/98	15	7.6	19.7	17.9	2.3	–72	261	139
8/21/98	17	7.7	25.7	19.9	[a]nm	–22	252	133
8/26/98	29	[b]—	—	—	—	160	(not calculated)	

Source: Data from Sutcliffe, K. D. 1999. *Dynamics of Irrigation-Induced and Saline Wet Soils, Central Utah*. MS thesis, Utah State University, Logan, UT.

Note: Water table depth, pH, EC, temperature, and DO averaged from measurements in duplicate 30-cm piezometers. Redox potential (Eh) represents the average of four redox probes. Anaerobic Eh threshold (the Eh below the soil is considered to be anaerobic) was calculated using pe = 12 pH, where pe = Eh/59 (McBride 1994). Technical Standard limit is the Eh value required for hydric soils by the USDA Hydric Soil Technical Standard. Eh values below the limit indicate (when soils are saturated) that the soils meet hydric soil conditions on that day.

[a] nm, not measured due to equipment failure.

[b] —, data not collected due to an insufficient amount of solution in the 30-cm piezometers.

In light of these data, one could argue that the presence of a salt crust might be used, at least partially, as a field indicator of a hydric soil subject to periodic saturation of the upper part of the soil from a saline water table. Ponded Aquisalids develop brittle salt crusts on the soil surface (Figure 15.5), whereas Aquisalids with soil saturation due to a capillary fringe develop fluffy, almost snow-like crusts (Boettinger 1997). The morphology of the crust, especially crystals, can be used to identify composition, precipitation history, and the nature of ponding during either annual or diurnal hydrologic activities (Last 1984).

A great amount of knowledge remains to be learned about the interaction between natural saline systems and redox chemistry. Additionally, many hydric soils in arid climates go undiscovered due to their poorly understood hydric soil morphology. As the importance of these areas rises due to demands in restoration, conservation, and preservation of sensitive and unique resources, opportunities for learning will also increase.

References

Arndt, J. L. and J. L. Richardson. 1989. Geochemistry of hydric soil salinity in a recharge–throughflow–discharge Prairie-Pothole wetland system. *Soil Sci. Soc. Am. J.* 53: 849–955.

Arndt, J. L. and J. L. Richardson. 1992. Carbonate and gypsum chemistry in saturated neutral pH soil environments. In R. D. Robarts, and M. L. Bothwell (Eds.) *Aquatic Ecosystems in Semi-Arid Regions: Implications for Resource Management.* NHRI Symposium Series 7, Environment Canada, Saskatoon, pp. 179–188.

Bai, J., H. Gao, W. Deng, Z. Yang, B. Cui, and R. Xiao. 2010. Nitrification potential of marsh soils from two natural saline–alkaline wetlands. *Biol. Fertil. Soils* 46: 525–529.

Boettinger, J. L. 1997. Aquisalids (Salorthids) and other wet saline and alkaline soils. Problems identifying aquic conditions and hydric soils. In M. J. Vepraskas, and S. Sprecher (Eds.) *Problems of Identifying Hydric Soils.* Soil Science Society of America Special Publication No. 50, Madison, WI, pp. 79–97.

Bolen, E. G., L. M. Smith, and H. L. Schramm Jr. 1989. Playa lakes: Prairie wetlands of the Southern High Plains. *BioScience* 39(9): 615–623.

Cariveau, A. B., D. C. Pavlacky, A. A. Bishop, and T. G. LaGrange. 2011. Effects of surrounding land use on playa inundation following intense rainfall. *Wetlands* 31: 65–73.

Eghbal, M. K., R. J. Southard, and L. D. Whittig. 1989. Dynamics of evaporate distribution in soils on a fan–playa transect in the Carrizo Plain, California. *Soil Sci. Soc. Am. J.* 53: 898–903.

Erickson, A. J. and V. L. Mortensen. 1974. *Soil Survey of Cache Valley Area, Utah: Parts of Cache and Box Elder Counties.* USDA Soil Conservation Service, U.S. Government Printing Office, Washington, DC.

Gile, L. H. and R. B. Grossman. 1979. *The Desert Project Soil Monograph.* U.S. Soil Conservation Service, U.S. Government Printing Office, Washington, DC.

Gumuzzio, J., J. Batille, and J. Casas. 1982. Mineralogical composition of salt efflorescences in a Typic Salorthid, Spain. *Geoderma* 28: 39–51.

Hardie, L. A. and H. P. Eugster. 1970. The evolution of closed-basin brines. *Miner. Soc. Am. Spec. Paper* 3: 273–290.

Heagle, D., M. Hayashi, and G. van der Kamp. 2013. Surface–subsurface salinity distribution and exchange in a closed-basin prairie wetland. *J. Hydrol.* 478: 1–14.

Keller, L. P., G. J. McCarthy, and J. L. Richardson. 1986. Mineralogy of soil evaporites in North Dakota. *Soil Sci. Soc. Am. J.* 50: 1069–1071.

Kohut, C., K. Muehlenbachs, and N. J. Dudas. 1995. Authigenic dolomite in a saline soil in Alberta, Canada. *Soil Sci. Soc. Am. J.* 59: 1499–1504.

Last, W. M. 1984. Sedimentology of playa lakes of the northern Great Plains. *Can. J. Earth Sci.* 21: 107–125.

Last, W. M. and P. DeDeckker. 1992. Paleohydrology and paleochemistry of Lake Beeac, a saline playa in southern Australia. In R. D. Robarts, and M. L. Bothwell (Eds.) *Aquatic Ecosystems in Semi-Arid Regions: Implications for Resource Management.* NHRI Symposium Series 7, Environment Canada, Saskatoon, pp. 63–74.

Lindsay, W. L. 1979. *Chemical Equilibria in Soils.* Wiley-Interscience, New York.

McBride, M. B. 1994. *Environmental Chemistry of Soils.* Oxford University Press, Oxford, UK.

Sherman, G. D., F. Schultz, and F. J. Alway. 1962. Dolomitization in soils of the Red River Valley, Minnesota. *Soil Sci.* 94: 304–313.

Skarie, R. L., J. L. Richardson, G. J. McCarthy, and A. Maianu. 1987. Evaporite mineralogy and groundwater chemistry associated with saline soils in eastern North Dakota. *Soil Sci. Soc. Am. J.* 51: 1372–1377.

Steinwand, A. L. and J. L. Richardson. 1989. Gypsum occurrence in soils on the margin of semi-permanent Prairie Pothole wetlands. *Soil Sci. Soc. Am. J.* 53: 836–842.

Sutcliffe, K. D. 1999. *Dynamics of Irrigation-Induced and Saline Wet Soils, Central Utah*. MS thesis, Utah State University, Logan, UT.

Timpson, M. E., J. L. Richardson, L. P. Keller, and G. J. McCarthy. 1986. Evaporite mineralogy associated with saline seeps in Southwestern North Dakota. *Soil Sci. Soc. Am. J.* 50: 490–493.

USDA NRCS. 2010. Field indicators of hydric soils in the United States, Version 7.0. In L.M. Vasilas, G.W. Hurt, and C.V. Noble (Eds.) *USDA, NRCS, in cooperation with the National Technical Committee for Hydric Soils*. http://www.nrcs.usda.gov/Internet/FSE_DOCUMENTS/nrcs142p2_053171.pdf.

U.S. Geologic Survey. 1970. *Selected Hydrologic Data, Cache Valley, Utah and Idaho*. Utah Basic Release Data No. 21, Salt Lake City, UT.

Whittig, L. D. and P. J. Janitzky. 1963. Mechanisms of formation of sodium carbonate in soils. I. Manifestations of biological conversions. *J. Soil Sci.* 14: 322–333.

Section III

Wetland Functions and Restoration

16

Wetland Soils and the Hydrogeomorphic Classification of Wetlands

Christopher V. Noble and Jacob F. Berkowitz

CONTENTS

Introduction

Purpose of Classification

The hydrogeomorphic (HGM) classification for wetlands developed as a starting point for applying functional assessments used in the determination of the effects of impacts to aquatic resources (Brinson 1993a, 1996). Classifying wetlands represents the first step in determining altered or degraded conditions relative to unaltered states. The HGM classification approach aggregates wetlands with similar geomorphic settings and hydrologic regimes. In so doing, wetland assessments can be tailored to a much narrower range of natural variation than if a single assessment procedure was designed for all wetland classes (Smith et al. 2013). Controlling for the degree of natural variation through classification allows for more efficient detection of alterations resulting from human activities (Smith et al. 1995). Although *Soil Taxonomy* (Soil Survey Staff 2010) is not addressed within the HGM classification approach, many of the factors associated with geomorphic setting and hydrologic regime relate to soils. In this chapter, we examine the application of wetland geomorphic setting and hydrologic regime (i.e., the HGM approach) to wetland soils.

The HGM classification considers three factors as critical to the functioning of wetlands: (1) geomorphic setting in the landscape, (2) dominant source of water, and (3) hydrodynamics (i.e., hydrologic regime; Brinson 1993a). We recognize that these three factors remain highly interdependent and autocorrelated, just as climate is highly influential on soil-forming processes in wetlands (Richardson 1997). The HGM classification approach identifies seven hydrogeomorphic groups: (1) riverine, (2) depressional, (3) slope, (4) mineral soil flats, (5) organic soil flats, (6) estuarine fringe, and (7) lacustrine fringe (Table 16.1). These classes differ from the Fish and Wildlife Service (FWS) wetland classification system outlined by Cowardin et al. (1979). The FWS National Wetland Inventory maps prove useful for wetland resource management, however, several nomenclature differences exist. For example, the HGM riverine class incorporates the entire river and its floodplain, while the FWS classification encompasses only the channel from bank to bank.

Position and movement of water in landscapes explain the distribution of wetlands, resulting in the separation of the landscape into uplands and aquatic environments, with wetlands occurring in transitional areas (Euliss et al. 2004; Mitsch and Gosselink 2007). Although the location of the boundary between wetlands and uplands has received considerable debate (NRC 1995; Tiner 1999), the boundary is really a part of a landscape continuum that is maintained by precipitation, which varies in both frequency and intensity (Brinson 1993b). Once a landscape receives precipitation, the water is redistributed until it is exported via stream flow, groundwater flow, or evapotranspiration (Carter 1986). In humid climates that support well-vegetated landscapes, the runoff factor is reduced to near zero. In warm, dry climates with low vegetation cover, evaporation becomes a more

TABLE 16.1

Hydrogeomorphic Classes of Wetlands Showing Associated Dominant Water Sources, Hydrodynamics, and Examples of Subclasses

			Examples of Subclasses	
Hydrogeomorphic Class	**Dominant Water Source**	**Dominant Hydrodynamics**	**Eastern U.S.**	**Western U.S. and Alaska**
Riverine	Over bank flow from channel	Unidirectional and horizontal	Bottomland hardwood forests	Riparian-forested wetlands
Depressional	Return flow from groundwater & interflow	Vertical	Prairie pothole marshes	California vernal pools
Slope	Return flow from groundwater	Uni-directional, horizontal	Fens	Avalanche chutes
Mineral soil flats	Precipitation	Vertical	Wet pine flatwoods	Large playas
Organic soil flats	Precipitation	Vertical	Peat bogs; portions of Everglades	Peat bogs
Estuarine fringe	Over bank flow from estuary	Bidirectional, horizontal	Chesapeake Bay marshes	San Francisco Bay marshes
Lacustrine fringe	Over bank flow from lake	Bidirectional, horizontal	Great Lakes marshes	Flathead Lake marshes

Source: Adapted from Brinson, M. M. et al. 1995. *Guidebook for Application of Hydrogeomorphic Assessments to Riverine Wetlands*. Technical Report TR-WRP-DE-11. Waterways Experiment Station, U.S. Army Corps of Engineers, Vicksburg, MS.

significant loss of water from the hydrologic cycle in contrast to more humid regions (Bullock and Acreman 2003).

The sections below describe characteristics associated with each HGM class illustrating: (1) geomorphic setting, (2) dominant hydrodynamic features, (3) at least one example of a soil hydrosequence, and (4) the most common field indicators of hydric soils encountered within each HGM class (NTCHS 2010). The HGM classes relate to wetland soil because geomorphic position and hydrologic processes impact soil-forming processes including genesis and morphology (Daniels and Gamble 1971; Richardson 1997). As such, soils reflect the long-term hydrology in wetlands (Tiner 1999). In addition to a discussion of the relationship between HGM classification and hydric soils, we introduce hydric soil measurements and characteristics commonly utilized in HGM functional assessments.

Description of HGM Classes

Depressional Wetlands

Geomorphic Setting

Depressional wetlands occur in basins that lie below the surrounding topography. Figure 16.1 depicts a basin containing a wet meadow in the center that is surrounded by a low prairie. This depression is typical of thousands in the prairie pothole glacial terrain (Stewart and Kantrud 1971).

Depressional wetlands have restricted surface outflow due to closed topographic contours (Millar 1976). Examples include interdunal areas in the sandhills of Nebraska, kettle

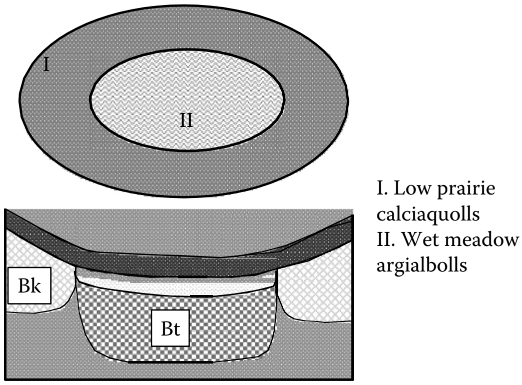


FIGURE 16.1
A seasonally ponded recharge wetland with wet meadow vegetation. The cross-section illustrates the soil distribution. (Adapted from Knuteson, J. A. et al. 1989. *Soil Sci. Soc. Am. J.* 53: 495–499.)

depressions in till, and some karst features such as sinkholes or dolines with high water tables (Kirkman et al. 2000). Some geographic regions are dominated by this class of wetland, such as the extensive glacial terrain in northern states, karst in the southeastern states, the various "playas" in Texas, and depressions of the intermountain region in the western U.S. (Bolen et al. 1989; Leibowitz and Vining 2003).

Arndt and Richardson (1989) and Steinwand and Richardson (1989) provide an illustration of till plain depressions in eastern North Dakota (Figure 16.2). These depressions remain completely closed to surface outflow in a subhumid, continental climate receiving 0.5 m annual precipitation with potential evapotranspiration of roughly 0.75 m. Salinity of the surface water varies greatly among wetlands as well as within a single wetland over time. The landscape is rolling ground moraine of relatively homogeneous till; the wetland density is high (often well over 80 depressions per square mile) and the surface is hummocky (Steinwand and Richardson 1989). Bedrock occurs 10 m or so below the lowest wetlands. The overlying till consists mainly of dead-ice facies of the Pleistocene Coleharbor Formation. The till contains dolomite and high sulfur marine shales (Arndt and Richardson 1989).

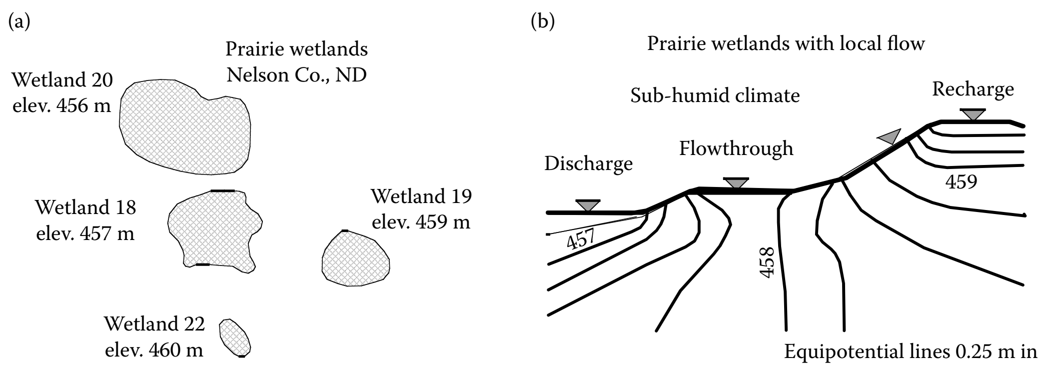


FIGURE 16.2
(a) Distribution of a few wetlands in the Arndt and Richardson (1989) study. (b) A flownet illustration of a general local flow based on a landscape in Nelson County, ND. (Adapted from Richardson, J. L., L. P. Wilding, and R. B. Daniels. 1992. *Geoderma* 53: 65–78.)

Depressional wetlands exhibit a wide variety of sizes and other characteristics, with plant communities occurring as described by Stewart and Kantrud (1971). For example Wetland 22 (Figure 16.2a) represents a small temporarily ponded area with wet meadow vegetation in the pond center (460 m above mean sea level); wetland 18 is a semipermanently ponded depression with deep marsh species in the center (457 m elevation); and wetland 20 is a larger saline lake with open water in the center (456 m). The lake dried out in 1988 (and probably in many other very dry years); the high salinity prevents establishment of emergent vascular plants.

Hydrology

Hydrogeomorphic classification aids in the determination of water movement in the landscape, and associated wetland processes (Shaffer et al. 1999; Merkey 2006). For example, the general flow of water in depressional wetland landscapes can be predicted based on landscape position and representative hydrologic observations (see Chapter 3). To illustrate this relationship, Figure 16.2b provides a cross-section of three depressions in North Dakota and illustrates equipotential lines (points of equal hydraulic head) in a depressional landscape. The water table is shown at the surface in the three wetlands, reflecting Sloan's (1972) comment, that "wetlands are windows to the water table." Flow of water is perpendicular to the equipotential lines. Under the depressional wetland marked recharge (wetland 22 in Figure 16.2a), the equipotential lines are roughly parallel to the surface and are decreasing. Water moves down and away from that wetland. In the flowthrough depression (wetland number 18 in Figure 16.2a), water intersects the wetland at the upper side (discharge) and recharges on the lower side. The equipotential lines are perpendicular to the land so that flow is through the wetland. In the lower depressional wetland, which is marked "Discharge" (wetland 20 in Figure 16.2a), the equipotential lines become more parallel and orient themselves with the surface. Because they are increasing upward, water is moving to the surface through the HGM classified depressional wetlands, investigators can begin to determine dominant patterns of hydrology including groundwater discharge and recharge.

Glacial landscapes form numerous and varied depressional wetlands (Gorham et al. 2007) and postglacial exposure of many depressional wetland landscapes creates groundwater flow systems modified by fracturing, local changes in stratigraphy, and soil types (Cook and Hauer 2007). Because of the numerous depressions in many till landscapes, the water flow is often in isolated local flow water systems (Toth 1963). In a subhumid climate, water transport occurs slowly from the upper ponds as depression-focused recharge. Depression-focused recharge, as used by Lissey (1971), implies that recharge in the center of the depression is the dominant hydrological process, although Hayashi et al. (1998) noted that most recharge water actually moves laterally and is lost to evaporation from the wetland edge. If a wetland center has both recharge and discharge, the wetland is a flowthrough wetland in which water discharges on one side and recharges on the other. This type of wetland is illustrated in the middle wetland in Figure 16.2b. In highly fractured tills, fewer flowthrough ponds occur because water is conducted downward away from nearby wetlands (Rosenberry and Winter 1997). In ablation till landscapes with typically coarser textured materials, flowthrough ponds occur frequently because buried and surficial aquifers are common. The lower wetlands receive water from the entire landscape as groundwater by a phenomenon known as depression-focused discharge. Interpond high areas participate in the flowthrough process less frequently in subhumid climates (Toth 1963). As water flows in long groundwater flow paths, wetlands in the low areas receive substantial amounts of dissolved ions and other solutes from the higher areas. The recharge wetlands release material, and the discharge wetlands accumulate the material.

In semiarid climates such as at St. Denis, Saskatchewan, studied by Miller et al. (1985), nearly all ponds (15 of 16) were recharge ponds. Water tables were mounded under the wetlands during wet periods and dropped quickly in dry periods. These observations are similar to the gaining or losing streams in the same climate. During drought cycles in subhumid areas, the drawdown of the water table, particularly at the edge of a wetland, creates a situation in which a pond may switch from being a discharge pond to being a recharge pond similar to the semiarid situation. Arndt and Richardson (1993) studied such a pond with a 20-year hydrologic record. The pond had always been a discharge pond and had a substantial amount of salt. During the 1988 drought, the pond became a recharge pond and dried up most of the year. It stayed dominantly dry for the next 3 years, during which time the salt was leached from the pond.

Each pond can also have a much smaller local flow system focused at the edge of the wetland. In a wetland studied by Whittig and Janitzky (1963), water ponded in the depression was evaporated from the edge, creating an accumulation in labile minerals illustrated by differentiated chemistry and soil. Their system was an edge-focused evaporative discharge type. Knuteson et al. (1989) observed a similar system (Figure 16.1) and measured the actual unsaturated or upward flow. From these estimates, they calculated that about 9000 years would be needed to form a Bk horizon. The flow envisioned by them is illustrated in Figure 16.3. Saturated flow in the pond interior moves materials down and away from the pond interior. This leaches the soil free of calcite and translocates the clay enough to form a Bt-horizon. The pond edges, however, remain an evaporative dry soil surface where wind and drying exert a tension on the wet soils below. This creates a water potential gradient such that matric tension moves the water from where the soil is wet to where it is drier. As the water evaporates, more water from the edge moves upward. The dissolved materials accumulate, and the edges become enriched with calcite, gypsum, and other salts and as illustrated in Figure 16.4. The B horizon at the edges is a Bkyzg: k means that substantial calcite has accumulated; y means that much gypsum has accumulated; z means that dissolved salts occur in the pores of the soil; and g means the soil is wet for substantial periods of time.

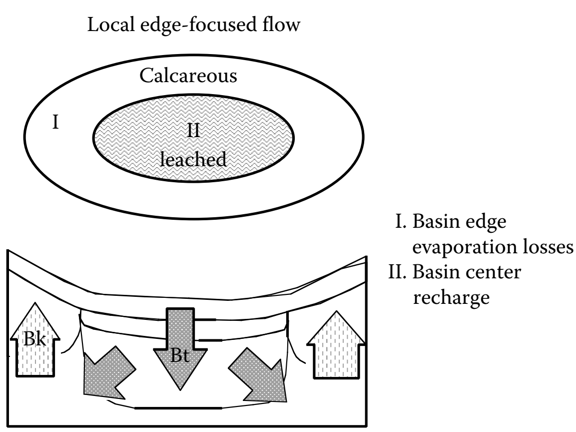


FIGURE 16.3
Flow in the pond center is saturated flow that recharges the water table and is dominantly downward and outward. Flow leaches the soils. The evapotranspiration on the edge of the pond creates an unsaturated upward flow with water loss. These soils will become enriched with calcite.

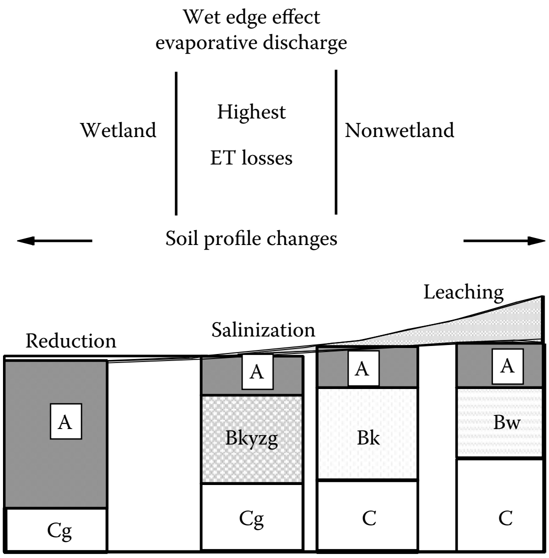


FIGURE 16.4
Wet soils at pond edges are subject to high evapotranspiration stresses because the water is near the surface and evaporation at the surface increases the matric tension and lifts more water to the surface. As the water is evaporated, the calcite increases, creating a Bk horizon.

Example Soil Hydrosequence

The HGM classification of depressional wetlands also provides insight into soil hydrosequences (Bell and Richardson 1997; Ballantine and Schneider 2009). The edge of a wetland, as indicated above, is the focus of evaporative discharge (Whittig and Janitzky 1963; Arndt and Richardson 1989; Knuteson et al. 1989; Steinwand and Richardson 1989; Seelig et al. 1990). For example, Figure 16.4 shows a wetland edge with two types of calcareous soils (Steinwand and Richardson 1989). This semipermanent flowthrough pond edge is similar to wetland 18 in Figure 16.2a. While the pond center has reducing conditions, the edge has strong evaporative discharge where a calcic (Bk) horizon forms due to the upward flow of water and a concentration of calcium. These two types of soils commonly occur in semipermanent ponds in the prairie areas: a wet soil that ranges from very poorly drained to poorly drained with a very gray Bk (Typic Calciaquoll) and a somewhat poorly drained soil with a khaki-colored Bk (Aeric Calciaquoll). The wetter soils frequently accumulate gypsum and dissolved salts (Steinwand and Richardson 1989), as indicated by the letters to designate the B horizon mentioned in the preceding section. The sequence here from driest to wettest is Typic Hapludolls, Aeric Calciaquolls, Typic Calciaquolls, and Cumulic Endoaquolls.

Field Indicators of Hydric Soils

Depending on the size and depth of the depression, soils vary greatly from the wetland edge to the deepest part of the wetland near the center. As a result, the hydroperiod of depressional wetlands ranges from only a few weeks of soil saturation at the depression edges to nearly permanent inundation of surface water near the center of the depression (Pyzoha et al. 2008). Hydric soil field indicators F3—Depleted Matrix, F6—Redox Dark Surface, and F8—Redox Depressions are commonly observed near the edge of depressional wetlands in fine-textured soils (see Chapters 8 and 9). Peat deposits may develop in the center of depressional wetlands that have zones with long periods of ponding during most years, leading to the development of A1—Histosol or A2—Histic Epipedon. In sandy soils S5—Sandy Redox and S6—Striped Matrix are commonly found at the depressional wetland edge. S4—Sandy Gleyed Matrix and F2—Loamy Gleyed Matrix are also common hydric soil indicators found in the deepest portion of depressional wetlands subject to extended periods of saturation or inundation. Several hydric soil field indicators are specific to depressional wetlands including F8—Redox Depressions, F9—Vernal Pools, F13—Umbric Surface, and F16—High Plains Depressions which remain limited for use in concave landforms.

Riverine Wetlands

Geomorphic Setting

Riverine wetlands include the HGM class occurring as linear features within the landscape that consist of a floodplain and stream channel (Vannote et al. 1980; Jurmu and Andrle 1997). Because riverine systems incise into the surrounding landscape, their topographically low position often creates discharge areas for groundwater (Winter 1999). Exceptions to this are "losing streams" of arid regions where the groundwater table slopes away from the channel and the floodplain (Vogt et al. 2010a,b). Riverine wetlands of floodplains of larger streams also receive water from upstream via overbank flow from the channel during flood events. The importance of this water source, relative to groundwater discharge, becomes greater as one moves downstream and the valley widens (Brinson 1993b).

The Wakarusa River valley near Lawrence, Kansas provides an example of riverine wetland hydrologic regime and soils. The Wakarusa entrenched its valley by cutting into an upland that is underlain by limestone bedrock. The escarpment that forms the boundary of the floodplain is a steep landform connecting the upland and floodplain (Figure 16.5). The floodplain has three landforms: (1) a back swamp, (2) a natural levee and entrenched channel of the Wakarusa River, (3) and a minor channel of a "yazoo" stream (Luft 1990). Back swamp, as used here, is a geomorphic term implying the area "in back of the natural levee" or away from the stream (Cazanacli and Smith 1998). Low, backwater areas receive water from the upland via the escarpment and flood water from the main stream channel via the natural levee (Figure 16.5). A smaller stream usually drains the back swamp and flows parallel to the trunk stream, such as the Yazoo River flowing parallel to the Mississippi River

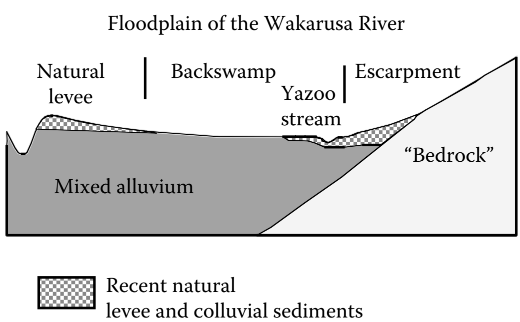


FIGURE 16.5
Cross-section of landforms for the Wakarusa River floodplain that includes the natural levee, back swamp, and the yazoo stream. These are bounded by the entrenched river channel and the escarpment.

in Mississippi; hence the name *yazoo* (Luft 1990). Yazoo streams often occur as straight drainageways traveling some distance parallel to the main stream. The backwater drainageways typically turn abruptly and re-enter the main channel (Wallerstein and Thorne 2004).

The back swamp riverine wetland observed in the Wakarusa River supports a variety of hydrophytes. In contrast, the elevated natural levee at this site supports non-hydrophytic tall prairie grass herbaceous species.

Hydrology

The dominant water sources for the HGM riverine wetland class includes overbank or backwater flow from the stream channel (Cole et al. 1997). Additional sources of water are direct precipitation, overland flow from adjacent uplands, and groundwater from subsurface connections to the main stream channel, hyporheic zone, or adjacent escarpment (Vannote et al. 1980). If the floodplain is cutoff from the stream channel by levees or severe downcutting, the riverine wetlands may function like flats with the primary water sources being precipitation (Brinson and Malvárez 2002). Within the Wakarusa River valley, both groundwater and surface water discharge from surrounding uplands into the back swamp wetlands where water becomes focused toward the yazoo. During most times of the year, the back swamp releases groundwater in the floodplain to the river by flowing laterally under the natural levee (Figure 16.6). During wetter times, such as during heavy rains, the natural levee acts as a drainage divide, increasing the water table within the back swamp riverine wetland (Figure 16.7). The slow throughflow system of water to the yazoo from the levee should be reflected in a progression of wetter soils toward the yazoo. The soils

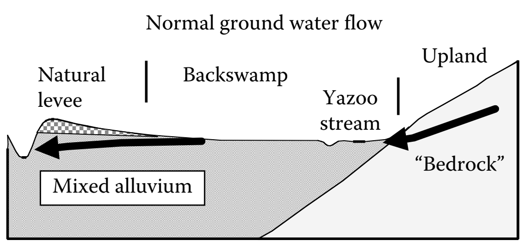


FIGURE 16.6
Flow reaches the river from the uplands by discharging on the floodplain near the escarpment. The yazoo stream carries away much of the water, but this stream is rather low gradient. Abundant water flows in the floodplain because of the permeable strata. Many floodplains are underlain by permeable strata.

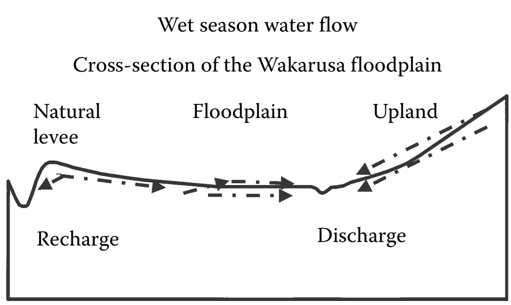


FIGURE 16.7
Natural levee acts as a drainage divide in wet periods, and the water table rises under the back swamp landform.

nearer the upland escarpment should be rather wet and possibly calcareous, reflecting the discharge of waters from the limestone that dominates the region. The transition of wetness and other soil characteristics (e.g., soil texture) from the back swamp to the escarpment would be quite sharp.

Example Soil Hydrosequence

The HGM classification of riverine wetlands provides insight into soil hydrosequences (Brooks et al. 2011, 2013). Within the Wakarusa River example, soil survey identifies two landform mapping units: the natural levee and the back swamp. The latter included the transitional soils that occurred on the lower levee (somewhat poorly drained and very poorly drained soils) and the soils down gradient from the escarpment. Typically, older soil surveys failed to differentiate poorly drained and very poorly drained soils in many areas, although the surveyors clearly were aware that these soils had the drainage inclusions. However, the distinction is now made due to the economic consequences of the Food Security Act, which requires more detailed wetland boundary identification. The soil mapping unit in the back swamp was divided into three units: a very poorly drained calcareous unit near the escarpment, a very poorly drained unit on both sides of the yazoo, and a nonhydric somewhat poorly drained to poorly drained unit transitional from the levee to the yazoo drainageway. Field observations of the vegetation reflected a strong correspondence between field indicators of hydric soil and hydrophytic vegetation (personal observations of the senior author and Kelly Kindscher of the Kansas Biological Survey, 1992).

Field Indicators of Hydric Soils

Wetland soils on floodplains exhibit extreme variability (Gallardo 2003; Bruland and Richardson 2005). Factors including stream gradient and discharge affect the amount and size of material transported within a stream and available to adjacent riverine wetlands and thus influence soil genesis and morphology. Additionally, deposited sediment transport varies as a function of soil texture, soil erodibility, and watershed land use (Fennessy et al. 1994; Arp and Cooper 2004). While not exclusively found in riverine wetlands, hydric soil field indicator A5—Stratified Layers is commonly found in flood plains. A5—Stratified layers is associated with areas that accumulate organic matter at the soil surface during wet, stable periods giving the soil a dark color. The surface layer then becomes buried by lighter colored stream sediments deposited during a flood event. Three field indicators of soils occur exclusively in flood plains associated with riverine wetlands: F12—Iron-Manganese Masses, F17—Delta Ochric, and F19—Piedmont Flood Plain Soils. Additionally, as observed in many HGM classes, F3—Depleted Matrix, F6—Redox Dark Surface, and S5—Sandy Redox remain common hydric soil field indicators across riverine wetlands.

Estuarine Fringe

Geomorphic Setting

Salt marshes are common HGM estuarine fringe coastal features found on nearly level landscapes behind barrier islands and spits, and along bays and lower tidal river shorelines (Wigland et al. 2012; Tiner 2013). Portions of these areas are flooded daily by tidal waters that carry abundant salts. Edmonds et al. (1985) describe the salt marsh soil types observed

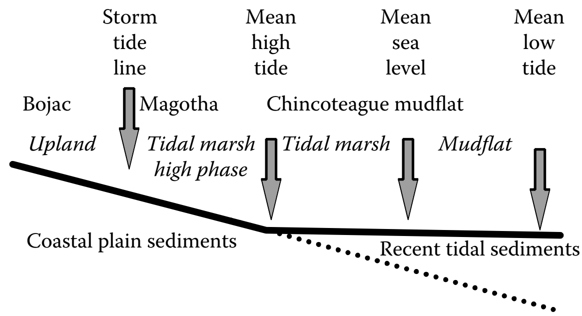


FIGURE 16.8
General landscape for Accomack and Northampton Counties, VA for the salt marsh landforms. (Adapted from Edmonds, W. J. et al. 1985. *Virginia Agric. Exp. Stat. Bull.* 85–88.)

and studied in Accomack and Northampton Counties of Virginia's eastern shore (37–38°N, 75–76°W). Their study encompasses 32,780 ha (81,000 acres) of seaside salt marshes that lie between the barrier islands and the mainland. The islands protect the marshes from storms in the Atlantic Ocean.

The Edmonds et al.'s (1985) study provides an example of a hydrogeomorphic unit for tidal systems. They separate their hydrogeomorphic unit into four soil-vegetation landform zones (Figure 16.8; Silberhorn and Harris 1977) including: (1) upland marine terraces with the Bojac series as a common representative which lies above the spring tide level; (2) salt meadow between storm tide line and the mean high tide represented by the Magotha series; (3) salt marsh cordgrass community between the mean high tide and mean sea level, represented by the Chincoteague series; and (4) tidal mudflats that lack vegetation of vascular plants. The Magotha soils represent former uplands located on the higher landscape positions in the salt marshes (Edmonds et al. 1985). In earlier soil surveys, landforms dominated by Magotha were a miscellaneous land class called Tidal Marsh, High Phase; the landforms occupied by Chincoteague soils were included in the miscellaneous land class Tidal Marsh.

Hydrology

The basic hydrologic feature of the estuarine fringe HGM class is the daily or intermittent tidal inundation by brackish water containing abundant sodium and other dissolved ions from sea water (Hughes et al. 1998) The frequency, depth, and duration of inundation varies considerably across topographic gradients, micro-, meso-, and macro-tidal regimes, as well as due to local climatic events such as storm surge (Callaway et al. 2012; Cooper 2013; Kirwan and Megonigal 2013). Within the coastal Virginia example illustrated in Figure 16.8, the interpretation of the hydrodynamics presented herein is based on the comments and data of Edmonds et al. (1985) and others at the Virginia Coast Reserve (Hmieleski 1994; Brinson et al. 1995; Stasavich 1998). The lower two landforms (tidal marsh and mud flat) flood and drain surficially, and thus remain saturated or have a peraquic moisture regime. The Chincoteague soils are dominated by saltwater cord grass (*Spartina alterniflora*), which is typically divided into tall, medium, and short grass growth forms. Tall forms tend to be restricted to creek bank environments, where flushing prevents accumulation of sulfides, and hypersaline conditions that tend to be associated with short growth forms (Delaune et al. 1983). Nitrogen supply may also be a factor (Broome et al. 1975). In the two lower landforms, the water table remains at or near the surface at low tide and above the surface at high tide.

The landform above the mean high tide and the spring tide line floods or saturates to the surface for extensive periods of time but precipitation typically provides the major source of water (Stasavich 1998). However, during storm and extreme tides, enough saltwater is transported to these sites to support the growth of marsh halophytes. Salt-meadow cord grass (*Spartina patens*) is the dominant plant, along with mixtures of coastal salt grass (*Distichlis spicata*) where drainage is more restricted and more saline. Roemer's rush (*Juncus roemerianus*) also occurs on this landform, although the study site occurs at the northern biogeographical distribution of the species.

In the lower tidal marsh and mud flat landforms, water table changes are surficial and only partly influenced by the soil itself. Typical weathering transformations that would cause profile development (other than reduction) do not take place because of the lack of infiltration and drawdown. Where coarser textured soils occur, such as on the barrier islands, groundwater discharge toward these landforms may stabilize the water table (Hayden et al. 1995). In fact, bioturbation by invertebrates, particularly fiddler crabs, and surficial sediment deposition are dominant factors in soil development. In the upper reaches of the Chincoteague landform, some drainage at low tide may allow for limited oxidation and minor translocation to a very shallow depth (Harvey et al. 1987), but most transport processes occur above rather than within the soil. In some locations, salt pans can develop where infrequent flooding (e.g., spring tides only), combined with evaporation, creates hypersaline conditions too salty for plant growth (Hayden et al. 1995).

In the high salt marsh areas, flow reversals of two types can be envisioned. The first is the spring tide flood or storm event bringing in saline waters. These events discharge water into the soils when unsaturated. Recession of the water and the subsequent lowering of the water table by evapotranspiration then allow precipitation and possibly some surface runoff from the upland to infiltrate. The landform maintains a relatively high water table in spite of infrequent flooding from estuarine sources (Stasavich 1998). Groundwater here is saline in contrast to fresh conditions in upland soils. In the transition between high marsh and forest, microrelief plays a role in the redistribution of salts. Microrelief highs act as local recharge areas, and the lows act as discharge areas, which flushes higher areas, resulting in lower soil salinity. Edmonds et al. (1985) mentioned that halophytes such as coastal salt grass (*Distichlis spicata*) and saltwort (*Salicornia* spp.) were present in lower areas, while trees such as loblolly pine (*Pinus taeda*) and eastern red-cedar (*Juniperus virginiana*) are restricted to hummocks. The combination of high water table and evaporation, called "evaporative discharge" by Seelig et al. (1990), is common in other saline landforms.

The uplands are freshwater-dominated systems that have better drainage than the salt marsh systems below them. These are freshwater recharge areas. The freshwater ponds above the denser saltwater, protecting the upland soils from encroachment of the saline water under the soils. In drier climates the saltwater table may move inland much farther because of the lower infiltration. The Bojac series, however, has a rapid infiltration and exists in an environment with annual precipitation that ranges from 25 to 60 in. (64–152 cm; Edmonds et al. 1985).

Example Soil Hydrosequence

Examining the soil hydrosequence (catena) for salt marshes provides insight into vegetation distributions, soil horizon development, and nutrient availability associated with

HGM estuarine fringe wetlands (Xiao et al. 2011; Twilley and Day 2012). For example, the salt marsh illustrated in Figure 16.9 depicts the relationship between landscape position, hydrologic regime, and soil development. The better-drained Bojac soil (non-hydric) is a coarse-loamy mixed thermic Typic Hapludult that is well drained and leached free of salts. Occurring between the spring tide line and the mean high tide, the Magotha series classifies as a coarse-loamy mixed thermic Typic Natraqualf. This soil is saline and sodic throughout its solum. If drained, the high sodicity of this soil would create a sodic condition resulting in a brick-hard consistency, due to the dispersion of clay caused by sodium. The tidal salt marsh soil, Chincoteague, reflects little profile development and classifies as fine-silty, mixed, nonacid thermic Typic Sulfaquent. If drained, it would become acidic as exposure and subsequent oxidation of sulfides results in sulfuric acid production (Burton et al. 2011) (Chapter 13). The mudflat is sediment.

The uplands contain leached soils with drier moisture regimes than the soils lower in the landscape (Figure 16.8). These soils create their own regional water table of freshwater. Soils in the salt marshes contain enough soluble salts that most are both saline and sodic (electrical conductivity >4.0 dS/m and sodium adsorption ratio >13) in sharp contrast to the adjacent upland. Sodium increases clay dispersion and possibly its translocation. In the lower tidal marsh and mud flat landforms, environmental conditions include high sodium and magnesium ion contents, chronic wetness, reducing conditions, accumulation of sulfate and chloride ions, and slow weathering. Accumulation of sulfide minerals in salt marsh soils results in acid sulfate soils with a drastic reduction of pH if these areas are drained and oxidized (Burton et al. 2011). For example, Edmonds et al. (1985) incubated Chincoteague soils in an oxidizing condition and measured a decrease from 7.0 to 3.0 in 24 days. The latter pH would significantly increase the solubility of aluminum, a plant toxin. The Magotha soil, however, did not significantly change in pH on incubation, which suggests it lacks sulfide accumulation.

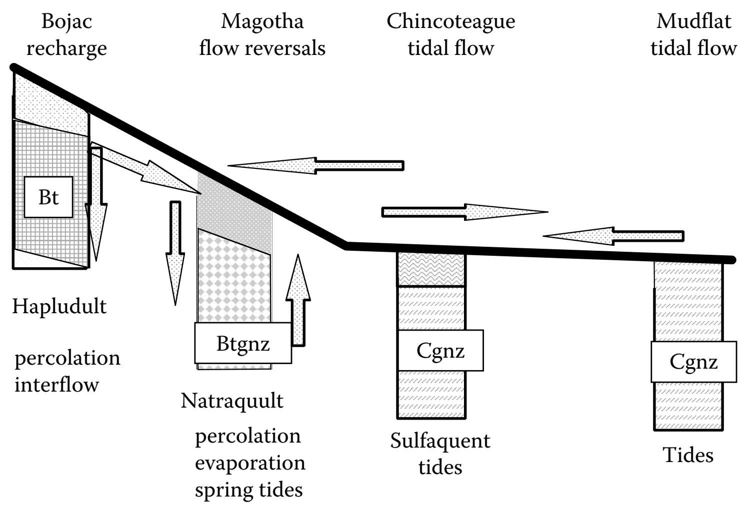


FIGURE 16.9
Soils schematic for salt marsh soils contrasted to upland and mudflat conditions. (Adapted from Edmonds, W. J. et al. 1985. *Virginia Agric. Exp. Stat. Bull.* 85–88.)

Field Indicators of Hydric Soils

Because of the daily tidal fluctuations observed within the estuarine fringe HGM class, tidal wetlands exhibit frequent saturation even in the absence of surface water inundation (Brooks et al. 2011). As a result, organic soil deposits (e.g., peat) occur within estuarine fringe wetlands as indicated by field indicators of hydric soil A1—Histosol or A2—Histic Epipedon. While not unique to estuarine fringe wetlands, hydric soil indicator A4—Hydrogen Sulfide is often found in the conditions of nearly constant periods of inundation and anaerobic conditions associated with daily tidal flooding and sea water-derived sulfur compounds. In sandy soils S5—Sandy Redox and S6—Stripped Matrix are commonly found at the wetland edge, while S4—Sandy Gleyed Matrix may occur down gradient. In fine-textured soils, F3—Depleted Matrix, F6—Redox Dark Surface, and F20—Anomalous Bright Loamy Soils are common hydric soil indicators near the edge and F2—Loamy Gleyed Matrix occurs in the wettest portion of estuarine fringes. Additionally, hydric soil field indicator F10—Marl occurs within estuarine fringe wetlands, including in the Florida Everglades where calcium rich HGM flat wetlands transition to coastal fringe systems.

Lacustrine Fringe

Geomorphic Setting

The lacustrine fringe HGM class includes wetlands associated with natural or impounded freshwater reservoirs (Brooks et al. 2011; Sueltenfuss et al. 2013) and unimpounded wetlands bordering large freshwater bodies (e.g., the Great Lakes; Cooper et al. 2013). For example, Figures 16.10a and 16.10b depict a lacustrine fringe wetland located on the western side of Lake Erie. The barrier sands create a lagoon system that extends from non-wetland down gradient to wet meadow, emergent marsh, and toward the open water of the lake. In this example, mineral soils dominate the wetlands at the soil surface. However, buried peat deposits also occur in the soil profile illustrating that water level fluctuations created and later destroyed fringe wetlands. The soil genesis sequence likely began with mineral wetland soils deposition, followed by Histosol development and subsequent burial during periods of high lake level and sedimentation. Currently, fringe wetlands along Lake Erie remain diked for waterfowl management impoundments. The dikes and causeways for roads and the canals in the wetlands and lagoons sever many of the original water connections with the lake.

Hydrology

The water level in the adjacent lake, reservoir, or impoundment provides the dominant water source for the lacustrine fringe HGM class wetlands. Precipitation, overland flow, and small groundwater discharges represent secondary water sources for lacustrine fringe wetlands. Dominant hydrodynamics is bidirectional from the water source into the wetland and return flow back to the water source. In many cases, lacustrine fringe systems intergrade with riverine wetlands where a river flows into a lake or estuary and flooding becomes the primary water source.

Example Soil Hydrosequence

Understanding the characteristics associated with the lacustrine fringe HGM class aids in the interpretation of soil–landscape interactions and ecology occurring in freshwater fringe wetlands (Wilcox 2012; Carling et al. 2013). For example, Figures 16.10a and 16.10b

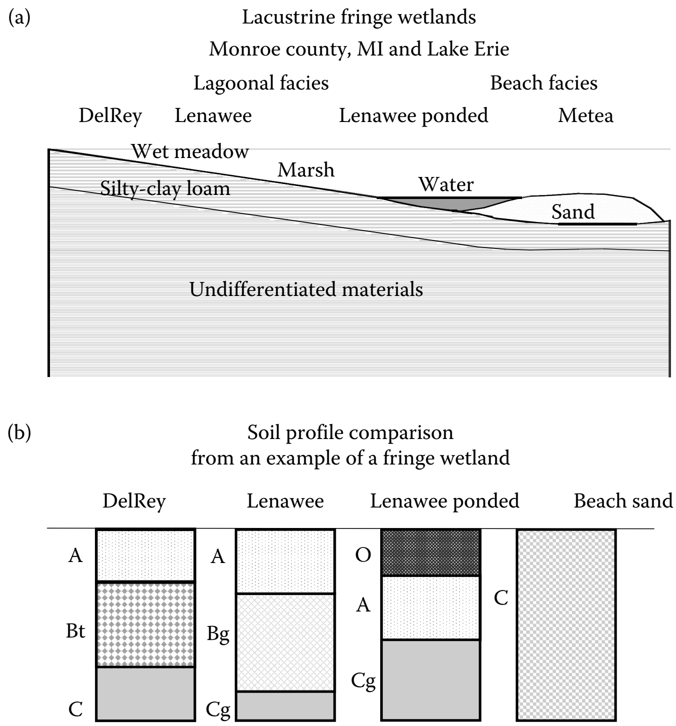


FIGURE 16.10
(a) Lacustrine fringe wetland based on a site along the western side of Lake Erie. (b) Example soil hydrosequence of lacustrine fringe soils in Monroe County, Michigan Soil Survey. (Adapted from Bowman, W. 1981. *Soil Survey of Monroe County, Michigan.* U.S. Govt. Printing Office, Washington, DC.)

depict an example of the soils and landforms associated with lacustrine fringe wetlands located on the western side of Lake Erie (Bowman 1981). The barrier sands yield to finer textured, wetter soils shoreward. As the water becomes shallower toward the upland, a marsh develops. The soil is mapped as Lenawee, fine, mixed, nonacid, mesic Mollic Epiaqepts. This high clay soil with a thin dark surface and neutral reaction is formed under conditions of "endo-saturation" or groundwater saturation. These soils occur in both the ponded marsh phase and the wet meadow phase, suggesting that two distinct soil taxa exist but are not separated. Inclusions of Saprists in the ponded marsh phase are high and may dominate some areas. The wet meadow phase can be farmed with some land modification. Herdendorf et al. (1981) relate the hydrophytes of these two mapping units. The somewhat poorly drained Del Rey series completes the hydrosequence. This Aeric Epiaqualf is fine textured with profile development suggesting frequent drying as well as ponding phases. The presence of carbonates within 2 or 3 feet of the surface and an argillic horizon indicate greater soil development than for the Lenawee, which is an Inceptisol lacking horizon development. The Lenawee soil does not dry out enough to allow for the downward movement of clay necessary to create an argillic horizon.

Field Indicators of Hydric Soils

Soils in lacustrine fringe HGM class wetlands often have field indicators of hydric soil similar to those found associated within HGM depressional wetlands depending on the stability of the water table level and length of inundation or saturation. Field indicators F3—Depleted Matrix and F6—Redox Dark surface remain common near the edge of lacustrine fringe wetlands in fine-textured soils. Organic soil deposits occur in fringe wetlands that have zones with long periods of ponding during most years, resulting in the development of hydric soil field indicators A1—Histosol, A2—Histic Epipedon, A10—2 cm Muck, and others. Floating mats represent a unique organic soil associated with the lacustrine fringe HGM class. In sandy soils S5—Sandy Redox and S6—Stripped Matrix are commonly found at the lacustrine fringe wetland edge. S4—Sandy Gleyed Matrix and F2—Loamy Gleyed Matrix are also common field indicators of hydric soil found in the wettest portion of lacustrine fringe wetlands.

Flats

Geomorphic Setting

The flats HGM class of wetlands occur on nearly level landscapes. Flats remain wet because of the lack of topographic relief resulting in limited lateral drainage water off of the landscape (Brooks et al. 2013; Noble et al. 2013). The geomorphic position of flats can be deceiving because they can occur high in the local landscape in geomorphic positions generally not associated with wetland formation. However, flats wetlands can cover large geographic areas (Zoltai and Pollett 1983). These areas were often ancient lake or ocean beds, and many flats have been extensively drained for agriculture (Sun et al. 2001; Whigham et al. 2007). Flats wetlands remain separated from the slope HGM class, which receive some groundwater input, because flats wetlands lack closed topographic contours that promote the concentration of water from the surrounding landscape.

Flats wetlands are divided into two separate HGM classes based on soils. Organic Flats wetlands exhibit surface soil layers 8 in. or more thick or occur over shallow bedrock (Brooks et al. 2013). Organic Flats typically are saturated near the soil surface or inundated for very long periods during most years (Verry et al. 2011; Laamrani 2014). Mineral Flats wetlands may exhibit thin organic surface layers, but often display mineral soil layers near the soil surface. Mineral Flats may only be inundated during short periods following precipitation events and may only saturate near the soil surface seasonally for a few weeks during most years (Noble et al. 2013). Pine flatwoods with hydric soils are an example of mineral flats (Kreye et al. 2014) and portions of the Everglades, northern peatlands and bogs provide examples of organic flats wetlands (Whigham and Jordan 2003).

The Edina series (fine, smectitic, mesic, Vertic Argialbolls) from Wayne County in southern Iowa provides an example of the mineral flat HGM class. The Edina series occurs on flat upland summit covered with 3 m of loess. Below the loess is a paleosol developed in highly weathered till of exceedingly high clay content. The map view of the landscape depicted in Figure 16.11 illustrates the dendritic stream dissection typical of this landscape and the flat upland.

Hydrology

Precipitation provides the dominant water source for the flats HGM class. Flats wetlands should not be confused with slope wetlands, in which groundwater discharge provides the primary water source (Brinson and Malvárez 2002). In flats wetlands, vertical fluctuation provides the dominant hydrodynamics. Flats wetlands lose water by mainly through evapotranspiration, or seepage to groundwater.

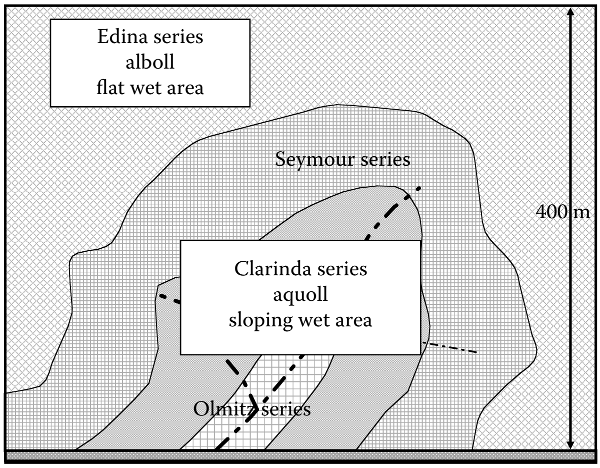


FIGURE 16.11
Distribution of soils on the landscape from the Wayne County, Iowa, Soil Survey. Note that the Edina series occurs in a nearly level landscape. Conversely, the other soil map units display 2%–7% slopes. (Adapted from Lockridge, L. D. 1971. *Soil Survey of Wayne County, Iowa*. USDA NRCS, U.S. Govt. Printing Office, Washington, DC.)

During the spring thaw in northern climates and rainy periods occurring in humid environments, the water on the landscape cannot run off easily because lateral flow is restricted by gradient rather than by texture. Downward movement is also often retarded by restrictive barriers. For example, the Edina series exhibits two aquitards capable of decreasing infiltration including the modern Bt horizon and the buried underlying paleosol (i.e., a soil that was buried and stopped forming) argillic horizon. The A and E horizons over the Bt horizon and the loess below the Bt horizon have relatively rapid permeability. The horizontal to downward saturated conductivity based on the NRCS estimated data is about 30/1. The combination of flat landscapes with low hydraulic gradient and restricted downward flow creates a large wet area. Figure 16.12 depicts the stratigraphy and flow in a flownet modeled after the Wayne County type location for Edina series (see Chapter 3 for background information). The flat is shown without any flow at all, though some may occur laterally in the thin soil surface. The perched water on top of the Bt horizon

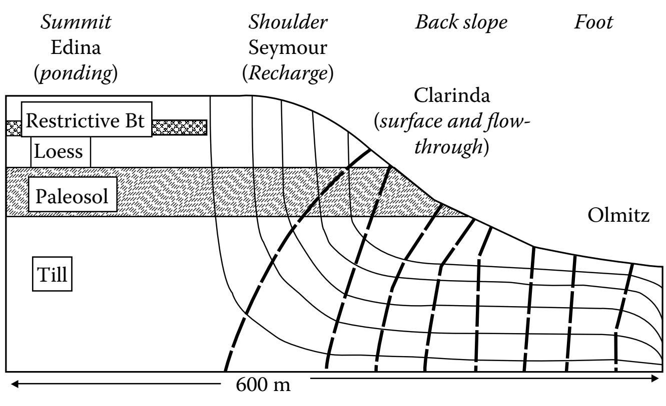


FIGURE 16.12
Flownet of the Edina landscape in the vicinity of Harvard, Wayne County, Iowa.

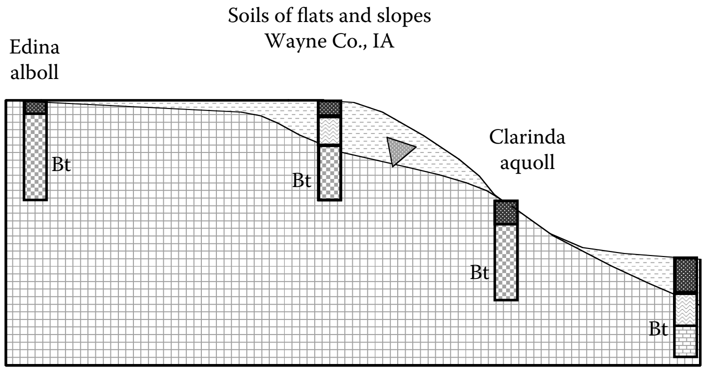


FIGURE 16.13
Cross-section with high water table and the soil types distributed on the landscape.

of the Alboll (Edina series) may saturate the horizons below, but the flow is so slow that the flowlines are concentrated in the shoulder position. This is the recharge area for flats (Richardson et al. 1992). A significant amount of water flows on top of the paleosol and discharges on the slope. The area used for this model includes a cove or headslope area (Figures 16.11 and 16.13). The convergence of flow in these areas creates a sloping wetland below the area occupied by the flats wetland.

The Edina example exemplifies the landscape position and hydrology associated with the HGM flats class and indicates that (1) flats wetlands become wet very fast and display slow lateral flows due to the low elevation gradient; (2) at the back-slope where the paleosol soil crops out, another wet area occurs; (3) recharge is concentrated at the shoulder positions; and (4) the flat releases little water to downward flow (i.e., infiltration). The Aquoll area developed on the paleosol is especially expressed in the coves or swales because of the convergence of lateral flowing water. The stratigraphy here produces potentially two HGM wetland classes, a flat and a sloping wetland. Local farmers are well aware of these wet areas because crops do not do well and tractors may get mired. The local name for these areas is "blue clays," and they are not spoken of with much fondness.

Example Soil Hydrosequence

The flat area of the landscape has a two-soil system. The interior of the flat area is wet and has an Alboll. The edge of the summit area has a better drained non-hydric soil (Figure 16.12). Daniels and Gamble (1967) called this the *red edge* after the reddish-colored soils in North Carolina in similar landscapes. These soils are located high on the landscape and therefore dry out late in the season. They are subject to translocation of clay and leaching of soluble constituents and develop a distinct profile. These are some of the few soils developed under prairie vegetation that have E or eluvial horizons reflective of the wetting and drying aspect of the soil.

Field Indicators of Hydric Soils

Soils in the flats HGM class wetlands often display a shallow restrictive layer (or even bedrock) retarding infiltration and perching water. In cold climates, permafrost or seasonal frost forms a layer that restricts water movement forming flats wetlands. Because precipitation provides the primary water source in flats wetlands, the HGM class often exhibits low nutrient concentrations and more acidity than slope wetlands. Soil development also depends strongly on the length of hydroperiod. Accumulation of organic matter at the soil surface is common in organic flat wetlands due to long periods of soil saturation and/or cold temperatures. Hydric soil field indications A1—Histosols, A2—Histic Epipedon, and A3—Black Histic are common in organic flats wetlands. Hydric soil field indications S2—Sandy Gleyed Matrix, F2—Loamy Gleyed Matrix, and A12—Thick Dark Surface are common hydric soil indicators in mineral flats soils subject to extensive periods of saturation, while F3—Depleted Matrix, F6—Redox Dark Surface, and A11—Depleted Below Dark Surface occur on seasonally flooded mineral flat wetlands. F10—Marl is a hydric soil indicator associated with Flats wetlands in portions of the Everglades that typically exhibit water table declines below the soil surface during the dry season.

Slope Wetlands

Geomorphic Setting

Slope wetlands are commonly referred to as fens, seeps, springs, carrs, headwaters, headslopes, bayheads, hollows, coves, and bay galls (Schafale and Weakley 1990). Slope wetlands occur where either (1) topographic position within the landscape or (2) geological conditions result in the discharge of groundwater into the wetland (Stein et al. 2004; Woods et al. 2006; Cole et al. 2008). As a result, these HGM subclasses have been defined as topographic slope wetlands and stratigraphic slope wetlands, respectively. The first relates to slopes that converge water in coves or draws. The second relates to stratum that intersects the land surface and forces the water to discharge onto the slope. In places, combinations of the two occur which amplifies discharge on slopes. The topographic slope wetland rarely occurs in semiarid and arid regions, but the stratigraphic type forms in any climate.

The topographic slope wetlands occur in concave convergent positions on landscapes, as illustrated in Figure 16.14, which shows the seasonally high water table position. Hack and Goodlett (1960) discussed the formation of these wet areas, which they called *hollows*, in the mountains of Virginia (other terms are *headslopes* and *coves*).

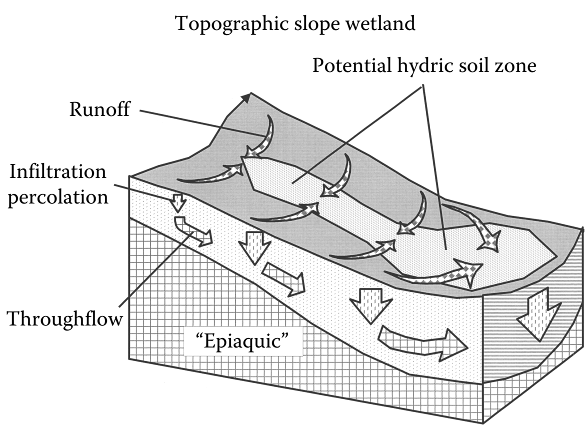


FIGURE 16.14
An illustration of a Topographic Slope Wetland with both runoff and throughflow water converging in the swale, creating an episaturated transient wetland.

Hydrology

Groundwater or interflow discharging at the land surface provides the dominant water source in slope wetlands, with direct precipitation contributing as a secondary source. Slope wetlands can be confused with flats wetlands, but in flats wetlands precipitation represents the primary water source. In slope wetlands, the dominant hydrodynamics occur downslope as unidirectional flow at or just below the soil surface. Slope wetlands lose water mainly through surface flow, shallow subsurface flow, and evapotranspiration. The convergence of flows occurs in zones at the margins of incipient channels that receive water from more than one direction. Thick soil provides the capacity to store water for long periods so that sudden rainfall events are followed by infiltration and slow movement in the landscape. The accumulation of the water at slope bases was noticeable to Hack and Goodlett (1960) and others from many landscapes (Chorley 1978). Areas of substantial wetness occur at the heads of drainages with short slopes and flat convergent zones with deep soils. Throughflow water moving by gravity is greatly slowed while infiltrating and moving in the soil. Penetration to depth in forest soils is often constricted by the soil subsurface horizons, such as argillic horizons, or from lack of macropores in the C horizon. Flow within the soil is slow if contrasted to runoff. However, once the pores are water filled, the wet area in the convergent landform expands upslope in all directions. The wettest area is the lower and central part of the convergent landform. Usually all soils in these landscapes are recharge zones that display signs of leaching.

These wet areas relate to the idea of "varying source area" of Hewlett and Nutter (1970). The wetlands that form expand up the slope with additional wetness. Nutter (1973) observed during his studies in the forests of the southeastern U.S. that water fed to the water table during storm events came from water that had been infiltrated and not from overland flow. Second, the water came not just from above a point on the landscape but also laterally from upslope and converged on the lower segments of the slope. Effective storage in these portions of the landscape was reduced. At the beginning of the drainage cycle actual flow may have been downward, but the net flow was downslope. As drainage continued, the flow lines slowly oriented more parallel with the surface. The upper boundary of topographic slope wetlands often remains diffuse, making it difficult to map for wetland delineation, especially if contrasted with the stratigraphic type of slope wetland. These wetlands typically contain mineral soils at the top of the slope, while Histosols often occur downslope if the concavity receives sufficient wetness. In the Howard County, Iowa situation described in the following section, the Histosol occurred in the flat area below the sloping portion of the wetland (Figure 16.15).

Kirkham (1947) conducted a wetness survey on areas that did not drain well despite having tile drains on the Iowan erosion surface in northeastern Iowa. These areas were foot slopes and usually had convergent water flow. On close inspection and measurement with piezometers, he determined that flow differed by landscape position. The flow was in the soil and little runoff occurred, even though some of the study area was cultivated. The upper areas were distinctly recharge areas with downward pressures. The side slopes had horizontal flow (parallel to the slope), and the lower slope areas had upward artesian pressures and discharge.

Example Soil Hydrosequence

The Howard County Soil Survey Report (Buckner and Highland 1974) reveals that the soils used by Kirkham were strongly anisotropic, and the impact on water had been observed (Figure 9.15). The Lourdes mapping unit is described as occurring on convex ridges and was an acid-leached soil. After heavy rains or extended wet periods, the water perches on

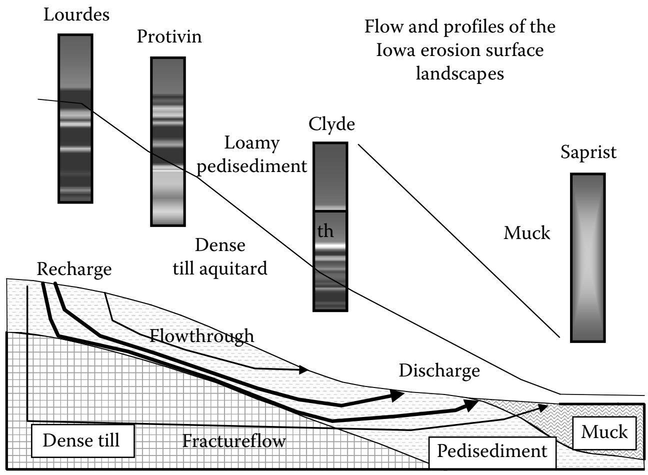


FIGURE 16.15
An area in Iowa with a topographic slope wetland that is tile drained. (Adapted from Kirkham, D. 1947. *Soil Sci. Soc. Am. Proc.* 12: 73–80; Buckner, W. and J. Highland. 1974. *Howard County Soil Survey Report.* U.S. Govt. Printing Office, Washington, DC.)

the impermeable dense lower till and creates side-hill seeps. Coupling the observations of Kirkham (1947) and later analyses (Nutter 1973; Chorley 1978), it seems that some deep water penetration occurs with abundant throughflow that discharges in the Clyde soil. The actual flow mechanism has created the downward flowing, well-drained Lourdes series exhibiting periodic wet periods with ponding. The water will flow laterally but is restricted by gradient and by saturated hydraulic conductivity. The sloping Protivin soil is deeper to the dense restrictive till stratum and receives water from above. This soil is somewhat poorly drained and has strong lateral flow tendencies. It is leached in its upper part but has carbonates in places in the restrictive stratum. The Clyde at the concave area of the slope is poorly drained and receives water from above. The soil of the flat area extending out from the hillslope wetland has a muck surface, which becomes deep enough to be a Saprist. This sequence is rather typical of fens; in fact Kratz et al. (1981) describe piezometric data in mounded peats similar to the sequence here but occurring almost entirely on Histosols.

Stratigraphic Slope Wetlands

Mausbach and Richardson (1994) described several aspects of fens, some of which are examples of stratigraphic slope wetlands. One example from Malterer et al. (1986) and Des Lauriers (1990) will be used here as an example. Stratigraphic slope wetlands occur because landscape geology creates exceptional anisotropic or directionally dependent conditions that focus water flow to a point on the landscape where the water discharges. Stratigraphic slope wetlands have sharp, narrow upper boundaries when contrasted to topographic slope wetlands. The strata conducting the water create a narrow transition area, just above the wetland boundary. Conversely, the diffuse nature of topographic slope wetland boundary displays a broad continuum of increasing wetness downslope. Figure 16.16 depicts a dense till with overlying sand and gravels of an outwash unit. The water moves freely in the gravels, but its downward movement is severely retarded in the till. The resulting point of discharge on the valley edge creates a calcareous fen with a 3% slope 15 m distance before starting to decrease to a nearly level contour. The soil types classify at the suborder level as

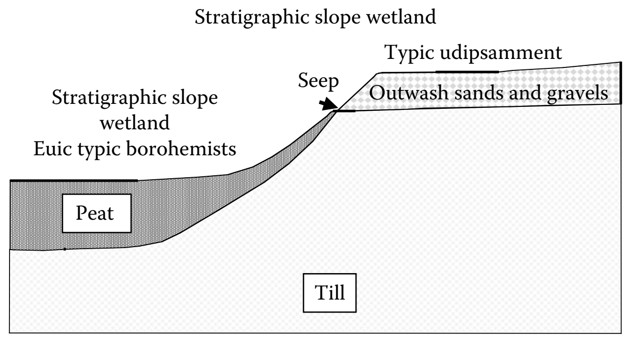


FIGURE 16.16
An illustration of a stratigraphic slope wetland that has developed into a fen with an organic soil; the area used to model this landscape is from the western part of North Dakota. (From Malterer, T. J. et al. 1986. *North Dakota Acad. Sci. Proc.* 40: 103.)

Hemists or Saprists (Malterer et al. 1986). The organic layer is >4 m thick at the base of the slope. The hydrology is simply that water discharges at the spring or seep on the hillslope. As the vegetation develops, some organic matter accumulates on the surface. The water tends to flow below the organic layer protected from evapotranspiration. The organic accumulation starts to act as an aquitard and confines the water to flow below the layer. The water often flows under positive head or artesian pressures as can be noted by the fountain created when the surface peaty-muck is penetrated with an auger or peat sampler. The water that moves through the landscape picks up substantial dissolved ions. These ions are concentrated and precipitated at the surface in places, but the high organic matter also holds the ions as adsorbed or exchangeable ions. The fens of stratigraphic slope wetlands remain nutrient rich compare with nutrient poor bogs that only receive rainwater. Notably, bogs would be considered in the HGM class of organic soil flats or depressions.

Field Indicators of Hydric Soils

Soils in slope wetlands tend to be calcareous (when carbonates occur in the landscape) because of the minerals transported to the soil surface by groundwater. Soil development is also strongly influenced by the length of hydroperiod which can be nearly all year because groundwater is often not solely provided by local precipitation. Soil temperature is also strongly influenced by the moderating influence of groundwater. As a result, accumulation of organic matter at the soil surface is common in slope wetlands due to long periods of soil saturation. A1—Histosols, A2—Histic Epipedon, and A3—Black Histic are common hydric soil indicators in slope wetlands. S2—Sandy Gleyed Matrix, F2—Loamy Gleyed Matrix, and A12—Thick Dark Surface are common hydric soil indicators in slope wetlands with mineral soils.

Application of Soil Characteristics to HGM Wetland Assessment

In addition to the HGM classification of wetland types described above, hydric soil characteristics also prove useful in interpreting and assessing the degree of disturbance or alteration in an area (Franklin et al. 2009; Smith et al. 2013). Many studies link disturbance and alteration to changes in hydric soil characteristics including soil texture, color, organic matter content, and others (Stolt et al. 2000; Bruland and Richardson 2005, 2006). As a result, changes in soil characteristics reflect wetland condition and the capacity to perform wetland functions. HGM examines components that can be measured rapidly during an onsite investigation (Smith et al. 2013). Further, HGM variables are intended to remain repeatable and capable of distinguishing between wetlands with minimal impacts and those exhibiting various levels and types of alteration or disturbance (Berkowitz et al. 2010). The requirements of rapid measurement and repeatability make soil characteristics including soil color, texture, organic matter accumulation, and horizon development ideal candidates for inclusion in HGM assessment methods. In the HGM approach, each variable measured is scored on a scale from 0.0 (indicating poor condition and function) to 1.0 (high condition or function) (Brinson, 1993). Scores are assigned relative to measurements collected in unaltered wetland locations (Smith et al. 1995). For example, if all undisturbed riverine wetlands within a study area display A horizons greater than 30 cm, and disturbed riverine wetlands consistently exhibit A horizons less than 30 cm thick, the thickness of the A horizon provides an indication of disturbance or alteration (Smith and Klimas 2002; Berkowitz 2013). The following section introduces several soil characteristics and variables used in one or more HGM wetland assessment methods.

Soil Detritus and O Horizon

The accumulation of soil detritus in wetlands provides a measure of organic matter input, processing, and storage; supplying energy subsidies to the food web and cover for soil invertebrates and other fauna (e.g., salamanders; Vannote et al. 1980; Meyer et al. 1998; Jung et al. 2004). For example, a number of studies demonstrate a decrease in detrital materials following disturbances such as agricultural clearing and logging (Yanai et al. 2003; Sun et al. 2004). As a result, measurements of detrital cover provide insight into wetland condition and function. Additionally, Berkowitz et al. (2010) demonstrated that measurements of detritus remain rapid and repeatable. The determination of soil detritus consists of measuring the percentage cover of detrital material on the soil surface. Soil detrital material is defined as the soil layer dominated by partially decomposed but still recognizable organic material, such as leaves, sticks, needles, flowers, fruits, insect frass, dead moss, or detached lichens on the surface of the ground (Soil Survey Staff 1993). Detrital material would classify as fibric or hemic material (peat or mucky peat). Detritus is a direct indication of short-term (1 or 2 years) accumulation of organic matter. Several HGM assessments incorporate soil detritus measurements (Smith and Klimas 2002; Noble et al. 2010; others). Reinhardt et al. (1997, 1999) and others also utilize measurements of soil detritus, however they refer to the parameter as "litter cover." The development of O horizons in wetlands indicates an accumulation of carbon at the soil surface that results from inundation and saturation and the onset of anaerobic conditions (Fanning and Fanning 1989; Reddy and DeLaune 2008).

A Horizon Thickness

The A horizons incorporate decomposed organic and mineral materials in near surface layers. Smith and Kimas (2002) indicate that determinations of the A horizon thickness provide a proxy measure for organic matter in wetland soils, defining the A horizon as the mineral soil horizon occurring below the O soil horizon, that consists of an accumulation of unrecognizable decomposed organic matter mixed with mineral soil. As seen in the soil detritus example above, O horizon and A horizon thickness reflect wetland disturbances, with intact hydric soils occurring in mature wetlands displaying more horizon development (Berkowitz 2013). Ainslie et al. (1999), Klimas et al. (2006), and others incorporate measurements of A horizon thickness into HGM wetland assessment methods.

Soil Color

In addition to measures of horizon development, HGM methods also examine soil color as an indicator of hydric soil disturbance. Hydric soils experience extended periods of inundation of saturation, resulting in the onset of anaerobic conditions, accumulation of organic material, and the reduction and translocation of iron and manganese oxides (Vepraskas and Sprecher 1997; Vepraskas 2004). These factors result in soil exhibiting low Munsell color value and chroma (Fanning and Fanning 1989). Soil organic matter accumulation accounts for low soil Munsell color values in wetlands, indicating long-term (at least several years) microbial decomposition of the detritus and incorporation into the soil. Direct measurement of the percentage of organic matter in the soil remains impractical for rapid HGM assessments. As a result, a relative determination of the soil organic matter content is made using soil color value. The scoring of soil value is based upon colors observed within limited HGM class and geographic area. For example, Noble et al. (2007) examined Munsell soil color values in slope wetlands located in Mississippi and Alabama. Munsell color values in undisturbed wetland areas remained below 2.0, while disturbed and altered hydric soils displayed soil color values as high as 7.0. Based on data from the least disturbed areas, a soil color score of 1.0 is assigned to wetland sites with average soil color values of 2.0 or less (Table 16.2). Munsell soil color values greater than 6.0 in the surface layer indicate a very low percentage of organic matter and severely altered conditions, resulting in a soil color score of 0.0. Intermediate soil color values (i.e., between 2.0 and 6.0) receive intermediate soil color scores (Table 16.2). Lee et al. (2001) and Noble et al. (2011) provide additional examples of the application of soil color values to HGM assessment methods.

Soil Texture

Common disturbances in hydric soils include filling and ditching (Dahl 2000; Bruland et al. 2003). Altering the texture of the soil through anthropogenic activities (e.g., fill, excavation,

TABLE 16.2

Surface Soil Color Score

Munsell Soil Color Value	Color Score
Less than or equal to 2	1.0
Greater than 2, but less than or equal to 3	0.8
Greater than 3, but less than or equal to 4	0.6
Greater than 4, but less than or equal to 5	0.4
Greater than 5, but less than or equal to 6	0.2
Greater than 6, but less than or equal to 10	0.0

Source: Adapted from Noble, C. V. et al. 2007. *Regional Guidebook for Applying the Hydrogeomorphic Approach to Assessing the Functions of Headwater Slope Wetlands on the Mississippi and Alabama Coastal Plains*. ERDC/EL TR-07-9. U.S. Army Engineer Research and Development Center, Vicksburg, MS.

rockplowing, cultivation) changes the capacity of water storage, cation exchange, impacts soil biological communities, and other factors (Bruland and Richardson 2005; Hartman et al. 2008). As a result, several HGM assessment methods incorporate determinations of soil textures as an indicator of disturbance. Noble et al. (2002) examined soil textures in marl wetlands occurring in the Florida Everglades and reported that unaltered areas contained muck or silty-textured soil materials, these sites received high soil texture scores (>0.9). Conversely, coarse surface textures (e.g., sands and gravels) and artificial surface (e.g., pavement) occurred in disturbed and altered soils, resulting in low soil texture scores (<0.2). Rheinhardt et al. (1997), Powell et al. (2003), and others also utilize measures of soil texture. In several cases, HGM assessment methods apply soil texture as a proxy measure for cation exchange capacity and soil integrity (Klimas and Smith 2002; Klimas et al. 2011).

Summary

Wetland soils are a key component of hydrogeomorphic classification of wetlands, as well as the development and application of wetland assessments. As outlined above, HGM combines geomorphic landscape settings and hydrologic features identifying seven distinct wetland classes that effect soil characteristics and processes. The HGM classes relate to wetland soil because geomorphic position and hydrologic processes impact soil-forming processes including genesis and morphology. As a result, HGM classification aids in the evaluation of pedogenic development, soil hydrosequence formation, and the determination of hydric soil functions. Further, we identified the most common field indicators of hydric soils associated with each HGM class. In addition to a discussion of the relationship between HGM classification and hydric soils, we introduced hydric soil measurements and characteristics commonly utilized in HGM functional assessments.

References

Ainslie, W. B., R. D. Smith, B. A. Pruitt, T. H. Roberts, E. J. Sparks, L. West, G. L. Godshalk. et al. 1999. *A Regional Guidebook for Assessing the Functions of Low Gradient, Riverine Wetlands in Western Kentucky*. Technical Report WRP-DE-17. U.S. Army Corps of Engineers, Waterways Experiment Station, Vicksburg, MI.

Arndt, J. L. and J. L. Richardson. 1989. Geochemical development of hydric soil salinity in a North Dakota prairie-pothole wetland system. *Soil Sci. Soc. Am. J.* 53: 848–855.

Arndt, J. L. and J. L. Richardson. 1993. Temporal variations in the salinity of shallow groundwater from the periphery of some North Dakota wetlands (USA). *J. Hydrol.* 141: 75–105.

Arp, C. D. and D. J. Cooper. 2004. Analysis of sediment retention in western riverine wetlands: The Yampa River watershed, Colorado, USA. *Environ. Manag.* 33(3): 318–330.

Ballantine, K. and R. Schneider. 2009. Fifty-five years of soil development in restored freshwater depressional wetlands. *Ecol. Appl.* 19(6): 1467–1480.

Bell, J. C. and J. L. Richardson. 1997 *Aquic Conditions and Hydric Soil Indicators for Aquolls and Albolls. Aquic Conditions and Hydric Soils: The Problem Soils*. Soil Society of America, Inc., Madison, WI, pp. 23–40.

Berkowitz, J. F., A. F. Casper, and C. V. Noble. 2010. A multiple watershed field test of hydrogeomorphic functional assessment of headwater streams—Variability in field measurements between independent teams. *Ecol.Indic.* 11(5): 1472–1475.

Berkowitz, J. F. 2013. Development of restoration trajectory metrics in reforested bottomland hardwood forests applying a rapid assessment approach. *Ecol. Indic.* 34: 600–606.

Bolen, E. G., L. M. Smith, and H. L. Schramm, Jr. 1989. Playa lakes: Prairie wetlands of the southern high plains. *BioScience* 39: 615–623.

Bowman, W. 1981. *Soil Survey of Monroe County, Michigan.* U.S. Govt. Printing Office, Washington, DC.

Brinson, M. M. 1993a. *A Hydrogeomorphic Classification for Wetlands*. Technical Report WRP-DE-4, Waterways Experiment Station, Army Corps of Engineers, Vicksburg, MS.

Brinson, M. M. 1993b. Gradients in the functioning of wetlands along environmental gradients. *Wetlands* 13: 65–74.

Brinson, M. M., F. R. Hauer, L. C. Lee, W. L. Nutter, R. D. Rheinhardt, R. D. Smith, and D. Whigham. 1995. *Guidebook for Application of Hydrogeomorphic Assessments to Riverine Wetlands*. Technical Report TR-WRP-DE-11. Waterways Experiment Station, U.S. Army Corps of Engineers, Vicksburg, MS.

Brinson, M. M. 1996. Assessing wetland functions using HGM. *Natl. Wetlands Newslett.* 18(1): 10–16.

Brinson, M. M. and A. I. Malvárez. 2002. Temperate freshwater wetlands: Types, status, and threats. *Environ. Conserv.* 29(02): 115–133.

Brooks, R. P., M. M. Brinson, K. J. Havens, C. S. Hershner, R. D. Rheinhardt, D. H. Wardrop, D. F. Whigham, A. D. Jacobs, J. M. Rubbo. 2011. Proposed hydrogeomorphic classification for wetlands of the mid-Atlantic region, USA. *Wetlands* 31(2): 207–219.

Brooks, R. P., M. M. Brinson, D. H. Wardrop, and J. A. Bishop. 2013. Hydrogeomorphic (HGM) classification, inventory, and reference wetlands. In *Mid-Atlantic Freshwater Wetlands: Advances in Wetlands Science, Management, Policy, and Practice*. Springer, New York, pp. 39–59.

Broome, S. W., W. W. Woodhouse, Jr., and E. D. Seneca. 1975. The relationship of mineral nutrients to growth of *Spartina alterniflora* in North Carolina. II. The effect of N, P, and Fe fertilizers. *Soil Sci. Soc. Am. J.* 39: 301–307.

Bruland, G. L., M. F. Hanchey, and C.J. Richardson. 2003. Effects of agriculture and wetland restoration on hydrology, soils, and water quality of a Carolina bay complex. *Wetlands Ecol. Manag.* 11(3): 141–156.

Bruland, G. L. and C. J. Richardson. 2005. Spatial variability of soil properties in created, restored, and paired natural wetlands. *Soil Sci. Soc. Am. J.* 69(1): 273–284.

Bruland, G. L. and Richardson, C. J. 2006. Comparison of soil organic matter in created, restored and paired natural wetlands in North Carolina. *Wetlands Ecol. Manag.* 14(3): 245–251.

Buckner, W. and J. Highland. 1974. *Howard County Soil Survey Report.* U.S. Govt. Printing Office, Washington, DC.

Bullock, A. and M. Acreman. 2003. The role of wetlands in the hydrological cycle. *Hydrol. Earth Syst. Sci.* 7: 358–389.

Burton, E. D., R. T. Bush, S. G. Johnston, L. A. Sullivan, and A. F. Keene. 2011. Sulfur biogeochemical cycling and novel Fe–S mineralization pathways in a tidally re-flooded wetland. *Geochim. Cosmochim. Acta* 75(12): 3434–3451.

Callaway, J. C., A. B. Borde, H. L. Diefenderfer, V. T. Parker, J. M. Rybczyk, and R. M. Thom. 2012. *Pacific Coast Tidal Wetlands* In D. P. Btazer and A. H. Baldwin (Eds.) University of California Press, Berkeley, CA, pp. 103–116.

Carling, G. T., D. C. Richards, H. Hoven, T. Miller, D. P. Fernandez, A. Rudd, E. Pazmino, W. P. Johnson et al. 2013. Relationships of surface water, pore water, and sediment chemistry in wetlands adjacent to Great Salt Lake, Utah, and potential impacts on plant community health. *Sci. Total Environ.* 443: 798–811.

Carter, V. 1986. An overview of the hydrologic concerns related to wetlands in the United States. *Can. J. Bot.* 64: 364–374.

Cazanacli, D. and N. D. Smith. 1998. A study of morphology and texture of natural levees—Cumberland Marshes, Saskatchewan, Canada. *Geomorphology* 25(1): 43–55.

Chorley, R. J. 1978. The hillslope hydrological cycle. In M. J. Kirkby (Ed.) *Hillslope Hydrology*. John Wiley & Sons, New York, NY, pp. 1–42.

Cole, C. A., R. P. Brooks, and D. H Wardrop. 1997. Wetland hydrology as a function of hydrogeomorphic (HGM) subclass. *Wetlands* 17(4): 456–467.

Cole, C. A., C. P. Cirmo, D. H. Wardrop, R. P. Brooks, and J. Peterson-Smith. 2008. Transferability of an HGM wetland classification scheme to a longitudinal gradient of the central Appalachian Mountains: Initial hydrological results. *Wetlands* 28(2): 439–449.

Cook, B. J. and F. R. Hauer. 2007. Effects of hydrologic connectivity on water chemistry, soils, and vegetation structure and function in an intermontane depressional wetland landscape. *Wetlands* 719–738.

Cooper, J. A. G. 2013. Mesoscale geomorphic change on low energy barrier islands in Chesapeake Bay, USA. *Geomorphology* 199: 82–94.

Cowardin, L. M., V. Carter, F. C. Golet, and E. T. LaRoe. 1979. *Classification of Wetland and Deepwater Habitats of the United States*. FWS/OBS-79/31. U.S. Fish and Wildlife Service, Washington, DC.

Dahl, T. E. 2000. *Status and Trends of Wetlands in the Conterminous United States 1986–1997*. US Fish and Wildlife Service, Washington, DC.

Daniels, R. B. and E. E. Gamble. 1967. The edge effect in some Ultisols in the North Carolina Coastal Plain. *Geoderma* 1: 117–124.

Daniels, R. B., E. E. Gamble, and L. A. Nelson. 1971. Relations between soil morphology and water-table levels on a dissected North Carolina coastal plain surface. *Soil Sci. Soc. Am. J.* 35(5): 781–784.

Darmody, R. G. and J. E. Foss. 1978. *Tidal Marsh Soils of Maryland*. Md. Agric. Exp. Stra. Misc. Publ, MD. p. 930.

DeLaune, R. D., C. J. Smith, and W. H. Patrick, Jr. 1983. Relation of marsh elevation, redox potential, and sulfide to Spartina alterniflora productivity. *Soil Sci. Soc. Am. J.* 47: 930–935.

Des Lauriers, L. L. 1990. *Soil Survey of McHenry County, North Dakota*. USDA Soil Conservation Service, U.S. Govt. Printing Office, Washington, DC.

Edmonds, W. J., G. M. Silberhorn, P. R. Cobb, C. D. Peacock, Jr., N. A. McLoda, and D. W. Smith. 1985. Soil classifications and floral relationships of seaside salt marsh soils in Accomack and Northampton counties, Virginia. *Virginia Agric. Exp. Stat. Bull*. 85–88.

Euliss, N. H., Jr., G. A. Swanson, J. W. LaBaugh, R. D. Nelson, T. C. Winter, D. O. Rosenberry, and D. M. Mushet. 2004. The wetland continuum: A conceptual framework for interpreting biological studies in the prairie pothole region. *Wetlands* 24: 448–458.

Fanning, D. E. and M. C. B Fanning. 1989. *Soil: Morphology, Genesis, and Classification*. John Wiley & Sons, New York.

Fennessy, S., C.C. Brueske, and W. J. Mitsch. 1994. Sediment deposition patterns in restored freshwater wetlands using sediment traps. *Ecol. Eng.* 3(4): 409–428.

Franklin, S. B., J. A. Kupfer, R. Pezeshki, R. Gentry, and R. D. Smith. 2009. Efficacy of the hydrogeomorphic model (HGM): A case study from western Tennessee. *Ecol. Indic.* 9(2): 267–283.

Gallardo, A. 2003. Spatial variability of soil properties in a floodplain forest in northwest Spain. *Ecosystems* 6(6): 564–576.

Gorham, E., C. Lehman, A. Dyke, J. Janssens, and L. Dyke. 2007. Temporal and spatial aspects of peatland initiation following deglaciation in North America. *Q. Sci. Rev.* 26(3): 300–311.

Groninger, J. W., W. M. Aust, M. Miwa, and J. A. Stanturf. 2000. Growth predictions for tree species planted on marginal soybean lands in the Lower Mississippi Valley. *J. Soil Water Conserv.* 55: 91–95.

Hack, J. T. and J. G. Goodlett. 1960. *Geomorphology and Forest Ecology of a Mountain Region in the Central Appalachians*. US Geol. Surv. Prof. Pap. No. 347. U.S. Govt. Printing Office, Washington, DC.

Hartman, W. H., C. J. Richardson, R. Vilgalys, and G. L. Bruland. 2008. Environmental and anthropogenic controls over bacterial communities in wetland soils. *Proc. Natl. Acad. Sci.* 105(46): 17842–17847.

Harvey, J. W., P. F. Germann, and W. E. Odum. 1987. Geomorphological control of subsurface hydrology in the creekbank zone of tidal marshes. *Estuarine Coast. Shelf Sci.* 25: 677–691.

Hayashi, M., G. van der Kamp, and D. L. Rudolph. 1998. Water and solute transfer between a prairie wetland and adjacent uplands, 1. Water balance. *J. Hydrol.* 207: 42–55.

Hayden, B. P., M. C. Rabenhorst, F. V. Santos, G. Shao, and R. C. Kockel. 1995. Geomorphic controls on coastal vegetation at the Virginia Coast Reserve. *Geomorphology* 13: 283–300.

Herdendorf, C. E., S. M. Hartley, and M. D. Barnes (Eds.) 1981. *Fish and Wildlife Resources of the Great Lakes Coastal Wetlands within the United States.* Vol. 1: *Overview.* FWS/OBS-81/02-v1. U.S. Fish and Wildlife Service, Washington, DC.

Hewlett, J. D. and W. L. Nutter. 1970. The varying source area of streamflow from upland basins. In *Proceedings of the Symposium on Interdisciplinary Aspects of Watershed Management.* Montana State Univ. Bozeman. Amer. Soc. Civil Engr, New York, NY, pp. 65–83.

Hmieleski, J. I. 1994. *High marsh-forest transitions in a brackish marsh: the effects of slope*. Master's thesis, Biology Department, East Carolina University, Greenville, NC.

Hughes, C. E., P. Binning, and G. R. Willgoose. 1998. Characterisation of the hydrology of an estuarine wetland. *J. Hydrol.* 211(1): 34–49.

Jung, R. E., P. Nanjappa, and H. C. Grant. 2004. *Stream Salamander Monitoring: Northeast Refuges and Parks. Northeast Amphibian Research and Monitoring Initiative*. Patuxent Wildlife Research Center, Laurel, MD.

Jurmu, M. C. and R. Andrle. 1997. Morphology of a wetland stream. *Environ. Manag.* 21(6): 921–941.

Kantrud, H. A., G. L. Krapu, and G. A. Swanson. 1989. *Prairie Basin Wetlands of the Dakotas: A Community Profile*. U.S. Fish and Wildlife Service, Washington, DC, USA, Biological Report 85(7.28).

Kirkham, D. 1947. Studies of hillslope seepage in the Iowan drift area. *Soil Sci. Soc. Am. Proc.* 12: 73–80.

Kirkman, L. K., P. C. Goebel, L. West, M. Drew, and B. Palik. 2000. Depressional wetland vegetation types: A question of plant community development. *Wetlands* 20(2): 373–385

Kirwan, M. L. and J. P. Megonigal. 2013. Tidal wetland stability in the face of human impacts and sea-level rise. *Nature* 504(7478): 53–60.

Klimas, C. V., E. O. Murray, H. Langston, T. Witsell, T. Foti, and R. Holbrook. 2006. *A Regional Guidebook for Conducting Functional Assessments of Wetland and Riparian Forests in the Ouachita Mountains and Crowley's Ridge Regions of Arkansas*. ERDC/EL TR-06-14. U.S. Army Engineer Research and Development Center, Vicksburg, MS.

Klimas, C. V., E. O. Murray, J. Pagan, H. Langston, and T. Foti. 2011. *A Regional Guidebook for Applying the Hydrogeomorphic Approach to Assessing Functions of Forested Wetlands in the Delta Region of Arkansas, Lower Mississippi River Alluvial Valley, Version 2.0*. ERDC/EL TR-11-12. U.S. Army Engineer Research and Development Center, Vicksburg, MS.

Knuteson, J. A., J. L. Richardson, D. D. Patterson, and L. Prunty. 1989. Pedogenic carbonates in a Calciaquoll associated with a recharge wetland. *Soil Sci. Soc. Am. J.* 53: 495–499.

Kratz, T. K., M. J. Winkler, and C. B. De Witt. 1981. Hydrology and chronology of a pear mound in Dane County southern Wisconsin. *Wisc. Acad. Sci. Arts Lett.* 69: 37–45.

Kreye, J. K., L. N. Kobziar, and J. M. Camp. 2014. Immediate and short-term response of understory fuels following mechanical mastication in a pine flatwoods site of Florida, USA. *Forest Ecol. Manag.* 313: 340–354.

Laamrani, A., O. Valeria, N. Fenton, Y. Bergeron, and L. Z. Cheng. 2014. The role of mineral soil topography on the spatial distribution of organic layer thickness in a paludified boreal landscape. *Geoderma*, 221: 70–81.

Lee, L. C., P. L. Fiedler, S. R. Stewart, R. R. Curry, D. J. Partridge, J. A. Mason, E. M. Inlander, R. B. Almy, D. L. Aston, and M. E. Spencer. 2001. *Draft Guidebook for Reference Based Assessment of the Functions of Riverine Waters/Wetlands Ecosystems in the South Coast Region of Santa Barbara County, California*. Santa Barbara County Water Agency, CA.

Leibowitz, S. G. and K. C. Vining. 2003. Temporal connectivity in a prairie pothole complex. *Wetlands* 23: 13–25.

Lissey, A. 1971. Depression-focused transient groundwater flow patterns in Manitoba. *Geol. Assoc. Can. Spec. Pap.* 9: 333–341.

Lockridge, L. D. 1971. *Soil Survey of Wayne County, Iowa*. USDA NRCS. U.S. Govt. Printing Office, Washington, DC.

Luft, E. R. 1990. Use of microscale landforms to teach introductory physical geography: Planning a local field exercise. *J. Geogr.* 89(3): 130–133.

Malterer, T. J., J. L. Richardson, and A. L. Duxbury. 1986. Peatland soils associated with the Souris River, McHenry County, North Dakota. *North Dakota Acad. Sci. Proc.* 40: 103.

Mausbach, M. J. and J. L. Richardson. 1994. Biogeochemistry processes in hydric soil formation. In W. H. Patrick, Jr. (Ed.) *Current Topics in Wetland Biogeochemistry*. Vol. 1. pp. 68–127.

Merkey, D. H. 2006. Characterization of wetland hydrodynamics using HGM and subclassification methods in southeastern Michigan, USA. *Wetlands* 6: 358–367.

Meyer, J. L., J. B. Wallace, and S. L. Eggert. 1998. Leaf litter as a source of dissolved organic carbon in streams. *Ecosystems* 1: 9.

Millar, J. B. 1976. *Wetland Classification in Western Canada: A Guide to Marshes and Shallow Open Water Wetlands in the Grasslands and Parklands of the Prairie Provinces*. Canadian Wildlife Service, Ottawa, ON, Canada. Report Series No. 37.

Miller, J. J., D. F. Acton, and R. J. St. Arnaud. 1985. The effect of groundwater on soil formation in a morainal landscape in Saskatchewan. *Can. J. Soil Sci.* 65: 293–307.

Mills, J. G. and M. Zwarich. 1986. Transient groundwater flow surrounding a recharge slough in a till plain. *Can. J. Soil Sci.* 66: 121–134.

Mitsch, W. J. and J. G. Gosselink, 2007. *Wetlands,* 4th ed. John Wiley & Sons, New York, NY.

National Research Council (NRC). 1995. *Wetlands: Characteristics and Boundaries. Committee on Characterization of Wetlands*. National Academy of Sciences, Washington, DC.

National Technical Committee for Hydric Soils (NTCHS). 2010. *Field Indicators of Hydric Soils in the United States (Version 7.0)*. United States Department of Agriculture, Natural Resources Conservation Service.

Noble, C.V., J. F. Berkowitz, and J. Spence. 2010 *Operational Draft Regional Guidebook for the Functional Assessment of High-Gradient Ephemeral and Intermittent Headwater Streams in Western West Virginia and Eastern Kentucky*. ERDC/EL TR-10-11. U.S. Army Engineer Research and Development Center, Vicksburg, MS.

Noble, C. V., E. O. Murray, C. V. Klimas, and W. Ainslie. 2011. *Regional Guidebook for Applying the Hydrogeomorphic Approach to Assessing the Functions of Headwater Slope Wetlands on the South Carolina Coastal Plain*. ERDC/EL TR-11-11. U.S. Army Engineer Research and Development Center, Vicksburg, MS.

Noble, C. V., T. H. Roberts, K. L. Morgan, A. J. Hill, V. S. Neary, and R. W. Cripps. 2013. *Regional Guidebook for Applying the Hydrogeomorphic Approach to Assessing the Functions of Flat and Seasonally Inundated Depression Wetlands on the Highland Rim*. ERDC/EL TR-13-12. U.S. Army Engineer Research and Development Center, Vicksburg, MS.

Noble, C. V., J. S. Wakeley, T. H. Roberts, and C. Henderson. 2007. *Regional Guidebook for Applying the Hydrogeomorphic Approach to Assessing the Functions of Headwater Slope Wetlands on the Mississippi and Alabama Coastal Plains*. ERDC/EL TR-07-9. U.S. Army Engineer Research and Development Center, Vicksburg, MS. Nutter, W. L. 1973. *The Role of Soil Water in the Hydrologic Behavior of Upland Basins. Field Soil Water Regime*. Soil Science Soc. Amer. Madison, WI, pp. 181–193.

Peacock, C. D., Jr. and W. J. Edmonds. 1992. Supplemental data for soil survey of Accomack County, *Virginia Agric Exp. Stat. Bull*. 92–93, Blacksburg, VA.

Powell, J. E., D. V. D'Amore, R. Thompson, P. Huberth, B. Bigelow, M. T. Walter, and T. Brock. 2003. *Wetland Functional Assessment Guidebook, Operational Draft Guidebook for Assessing the Functions for Riverine and Slope River Proximal Wetlands in Coastal Southeast and Southcentral Alaska Using the HGM Approach*. State of Alaska Department of Environmental Conservation.

Pyzoha, J. E., T. J. Callahan, G. Sun, C. C. Trettin, and M. Miwa. 2008. A conceptual hydrologic model for a forested Carolina bay depressional wetland on the Coastal Plain of South Carolina, USA. *Hydrol. Process.* 22(14): 2689–2698.

Reddy, K. R. and R. D. DeLaune. 2008. *Biogeochemistry of Wetlands*. CRC Press, Boca Raton, FL.

Rheinhardt, R. D., M. M. Brinson, and P. M. Farley. 1997. Applying reference wetland data to functional assessment, mitigation and restoration. *Wetlands* 17: 195–215.

Rheinhardt, R. D., M. C. Rheinhardt, M. M. Brinson, and K. E. Faser, Jr. 1999. Application of reference data for assessing and restoring headwater ecosystems. *Restor. Ecol.* 7: 241–51.

Richardson, J. L., L. P. Wilding, and R. B. Daniels. 1992. Recharge and discharge of groundwater in aquic conditions illustrated with flownet analysis. *Geoderma* 53: 65–78.

Richardson, J. L. 1997. Soil development and morphology in relation to shallow ground water: An interpretation tool. In K. W. Watson and A. Zaporozec (Eds.) *Proceedings of the 4th Decade of Progress Symposium, Tampa Bay, FL*. American Institute of Hydrology, St. Paul, MN, pp. 229–233.

Rosenberry, D. O. and T. C. Winter. 1997. Dynamics of water-table fluctuations in a upland between two prairie-pothole wetlands in North Dakota. *J. Hydrol.* 191: 266–289.

Schafale, M. P. and Weakley, A. S. 1990. *Classification of the Natural Communities of North Carolina. Third Approximation*. North Carolina Department of Environment, Health, and Natural Resources, Division of Parks and Recreation, Natural Heritage Program, Raleigh.

Seelig, B. D., J. L. Richardson, and W. T. Barker. 1990. Characteristics and taxonomy of sodic soils as a function of landform position. *Soil Sci. Soc. Am. J.* 54: 1690–1697.

Shaffer, P. W., M. E. Kentula, and S. E. Gwin. 1999. Characterization of wetland hydrology using hydrogeomorphic classification. *Wetlands* 490–504.

Silberhorn, G. M. and A. F. Harris. 1977. *Accomack County Tidal Marsh Inventory*. Spec. Rep. No. 138. Applied Marine Science and Ocean Engineering, Virginia Institute Marine Science, Gloucester Point, VA.

Sloan, C. E. 1972. *Ground-Water Hydrology of Prairie Potholes in North Dakota*. U.S. Geol. Survey Prof. Pap. 585-C. U.S. Govt. Printing Office, Washington, DC.

Smith, R. D., A. Ammann, C. Bartoldus, and M. M. Brinson. 1995. *An Approach for Assessing Wetland Functions Using Hydrogeomorphic Classification, Reference Wetlands and Functional Indices*. Technical Report TR-WRP-DE-9. Waterways Experiment Station, U.S. Army Corps of Engineers, Vicksburg, MS.

Smith R. D. and C. V. Klimas. 2002. *A Regional Guidebook for Applying the Hydrogeomorphic Approach to Assessing Wetland Functions of Selected Regional Wetland Subclasses, Yazoo Basin, Lower Mississippi River Alluvial Valley*. ERDC/EL TR-02-4. U.S. Army Engineer Research and Development Center, Vicksburg, MS.

Smith, R. D., C. V. Noble, and J. F. Berkowitz, J. F. 2013. *Hydrogeomorphic (HGM) Approach to Assessing Wetland Functions: Guidelines for Developing Guidebooks (Version 2)*. ERDC/EL TR-13-11. U.S. Army Engineer Research and Development Center, Vicksburg, MS.

Soil Survey Staff. 2010. *Keys to Soil Taxonomy*, 11th ed. USDA-Natural Resources Conservation Service, Washington, DC.

Stasavich, L. E. 1998. *Quantitatively Defining Hydroperiod with Ecological Significance to Wetland Functions*. In progress. MS thesis, Biology Department, East Carolina University, Greenville, NC.

Stein, E. D., M. Mattson, A. E. Fetscher, and K. J. Halama. 2004. Influence of geologic setting on slope wetland hydrodynamics. *Wetlands* 24(2): 244–260.

Steinwand, A. L. and J. L. Richardson. 1989. Gypsum occurrence in soils on the margin of semipermanent prairie pothole wetlands. *Soil Sci. Soc. Am. J.* 53: 836–842.

Stewart, R. E. and H. C. Kantrud. 1971. Classification of natural ponds and lakes in the glaciated Prairie region. *U.S. Fish. Wild. Serv., Resour. Publ*. 92: 57.

Stolt, M. H., M. H. Genthner, W. L. Daniels, V. A. Groover, S. Nagle, and K. C. Haering. 2000. Comparison of soil and other environmental conditions in constructed and adjacent palustrine reference wetlands. *Wetlands* 20(4): 671–683.

Sueltenfuss, J. P., D. J. Cooper, R. L. Knight, and R. M. Waskom. 2013. The creation and maintenance of wetland ecosystems from irrigation canal and reservoir seepage in a semi-arid landscape. *Wetlands* 33(5): 799–810.

Sun, G., S. G. McNulty, J. P. Shepard, D. M. Amatya, H. Riekerk, N. B. Comerford, W. Skaggs, L. Swift. 2001. Effects of timber management on the hydrology of wetland forests in the southern United States. *Forest Ecology and Management* 143(1): 227–236.

Tiner, R. W. 1999. *Wetland Indicators: A Guide to Wetland Identification, Delineation, Classification, and Mapping*. CRC Press, Boca Raton, FL.

Tiner, R. W. 2013. *Tidal Wetlands Primer*. University of Massachusetts Press, MA.

Toth, J. 1963. A theoretical analysis of groundwater flow in small drainage basins. In *Proc. Hydrol. Symp. Groundwater* Vol. 3. Queen's Printer, Ottawa, Canada, pp. 75–96.

Twilley, R. R. and J. W. Day Jr. 2012. *Mangrove Wetlands. Estuarine Ecology*. Wiley, New York, pp. 165–202.

Vannote, R. L., G. W. Minshall, K. W. Cummins, J. R. Sedell, and C. E. Cushing. 1980. The river continuum concept. *Can.J. Fish. Aquat. Sci.* 37(1): 130–137.

Vepraskas, M. J. and S. W. Sprecher. 1997. Overview of aquic conditions and hydric soils. In M. J. Vepraskas and S. W. Sprecher (Eds.) *Aquic Conditions and Hydric Soils*. Soil Science Society of America, Madison, WI.

Vepraskas, M. J. 2004. *Redoximorphic Features for Identifying Aquic Conditions*. North Carolina Agricultural Research Service Technical Bulletin No. 301. North Carolina State University, CA.

Verry, E. S., D. H. Boelter, J. Paivanen, D. S. Nichols, T. Malterer, and A. Gafni. 2011. Physical properties of organic soils. *Peatland Biogeochemistry and Watershed Hydrology at the Marcell Experimental Forest*. CRC Press, Boca Raton, FL, 135–176.

Vogt, T., E. Hoehn, P. Schneider, A. Freund, M. Schirmer, and O. A. Cirpka. 2010a. Fluctuations of electrical conductivity as a natural tracer for bank filtration in a losing stream. *Adv. Water Resources* 33(11): 1296–1308.

Vogt, T., P. Schneider, L. Hahn-Woernle, and O. A. Cirpka. 2010b. Estimation of seepage rates in a losing stream by means of fiber-optic high-resolution vertical temperature profiling. *J. Hydrol.* 380(1): 154–164.

Wallerstein, N. P. and C. R. Thorne. 2004. Influence of large woody debris on morphological evolution of incised, sand-bed channels. *Geomorphology* 57(1): 53–73.

Whigham, D. F. and T.E. Jordan. 2003. Isolated wetlands and water quality. *Wetlands* 23(3): 541–549.

Whigham, D. F., A. D. Jacobs, D. E. Weller, T. E. Jordan, M. E. Kentula, S. F. Jensen, and D. L. Stevens. 2007. Combining HGM and EMAP procedures to assess wetlands at the watershed scale—Status of flats and non-tidal riverine wetlands in the Nanticoke River watershed, Delaware and Maryland (USA). *Wetlands* 27(3): 462–478.

Whittig, L. D. and P. Janitzky. 1963. Mechanisms of formation of sodium carbonate in soils. I. Manifestations of biological conversions. *J. Soil Sci*. 14: 322–333.

Wigland, C. and C. T Roman. 2012. North Atlantic coastal tidal wetlands. In D. P. Btazer and A. H. Baldwin (Eds.) *Wetland Habitats of North America*. University of California Press, Berkeley, CA, pp. 13–28.

Wilcox, D. A. 2012. Great Lakes coastal marshes. *Wetland Habitats of North America: Ecology and Conservation Concerns*. University of California Press, Berkeley, CA, pp. 173–188.

Winter, T. C. 1999. Relation of streams, lakes, and wetlands to groundwater flow systems. *Hydrogeol. J.* 7(1): 28–45.

Woods, S. W., L. H. MacDonald, and C. J. Westbrook. 2006. Hydrologic interactions between an alluvial fan and a slope wetland in the central Rocky Mountains, USA. *Wetlands* 26(1): 230–243.

Xiao, R., J. Bai, H. Zhang, H. Gao, X. Liu, and A. Wilkes. 2011. Changes of P, Ca, Al and Fe contents in fringe marshes along a pedogenic chronosequence in the Pearl River estuary, South China. *Continental Shelf Res.* 31(6): 739–747.

Yanai, R. D., W. S. Currie, and C. L. Goodale. 2003. Soil carbon dynamics after forest harvest: An ecosystem paradigm reconsidered. *Ecosystems* 6(3): 197–212.

Zoltai, S. C. and F. C. Pollett. 1983. *Wetlands in Canada, Their Classification, Distribution, and Use*. Elsevier Scientific, New York, NY, pp. 246–268.

17

Approaches to Assessing the Ecological Condition of Wetlands Using Soil Indicators

M. Siobhan Fennessy and Denise Wardrop

CONTENTS

Introduction

Wetlands occupy a relatively small portion of the earth's land surface, with estimates of their global extent ranging from 5% to 8%, or between 5.3 and 12.8 million km^2. Approximately half of that area has been lost and much of the area that remains is degraded due to human activities (Mitsch and Gosselink 2007). Wetlands provide a suite of ecological functions, generally classified as hydrologic, biogeochemical, or habitat support functions. When these are valued by society, they are generally referred to as ecosystem services and include flood water storage, carbon (C) sequestration, water quality improvements, and habitat provisioning. The realization that wetland ecosystem services are critical for human health and well-being (Millennium Ecosystem Assessment 2005) has illustrated the need for assessment protocols that can provide estimates of the level of service provided, detect the impact of human activities on their ecological condition, and guide us in restoration efforts (Zedler 2003). In this chapter, we review some of the key approaches that have been developed to employ soil characteristics in the assessment of both the ecological condition and the functional capacity of wetlands.

Wetlands are among the ecosystems most altered by human activities (Mitsch and Gosselink 2007). Disturbances such as hydrologic alterations, nutrient enrichment, and land use change generate stressors that lead to changes in ecosystem processes and the ecosystem services provided (Figure 17.1). In visualizing this as a dose–response relationship (Figure 17.2), ecological condition is expressed on the y-axis, and represents the extent to which a site departs from the full measure of ecological integrity. Ecological integrity is the ability of an ecosystem to support and maintain its complexity and capacity for

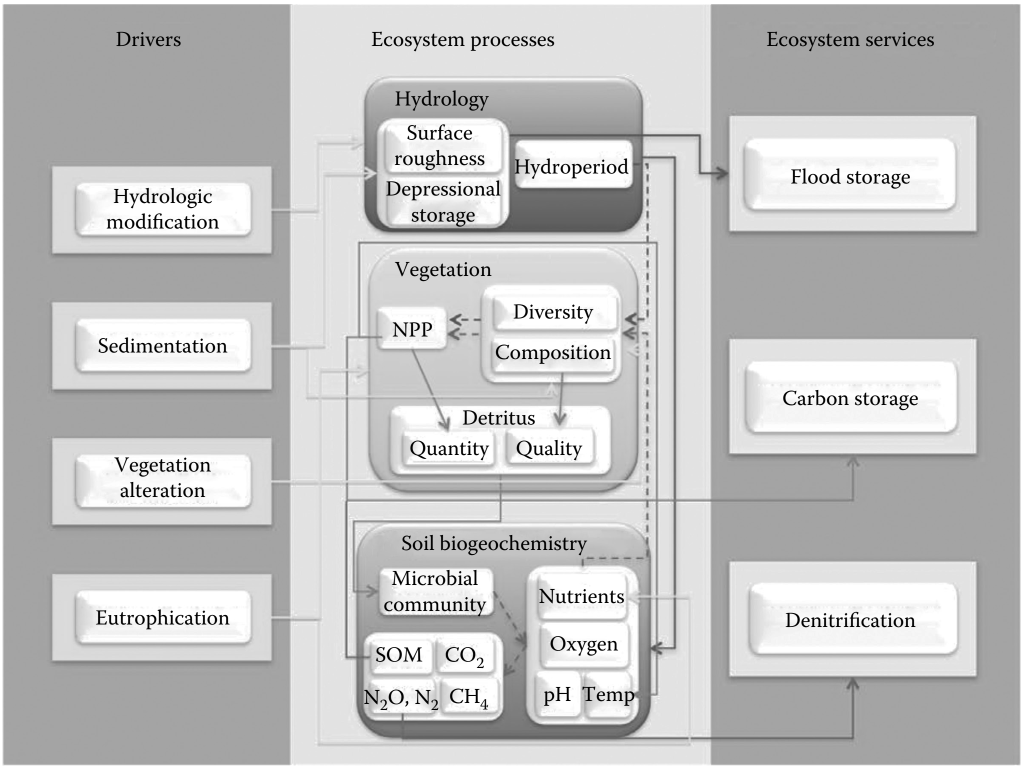


FIGURE 17.1
The relationship between the human drivers of ecosystem change, the response of ecosystem processes within wetlands, and the response as seen in alterations of ecosystem services. The model describes the relationships between landscape and ecosystem response through links between intermediate structural wetland components.

self-organization in terms of its physicochemical characteristics, species composition, and functional processes, in the absence of human disturbance (Karr and Dudley 1981). Measures of individual functions/ecosystem services can also be represented on the y-axis, instead of general condition. The x-axis is an expression of anthropogenic stress, and can be measured by either an individual stressor, such as sedimentation, or as a multistressor indicator, such as land cover change (Moon and Wardrop 2013).

The concept of soil quality naturally aligns with these definitions of ecological integrity and condition. For the purposes of this chapter, we use the definition of soil quality published by the Soil Science Society of America's Ad Hoc Committee on Soil Quality (S-581), "the capacity of a specific kind of soil to function, within natural or managed ecosystem boundaries, to sustain plant and animal productivity, maintain or enhance water and air quality, and support human health and habitation" (Karlen et al. 1997). Because this definition includes the identified suite of wetland functions (hydrologic, biogeochemical, and habitat), soil quality can be thought of as the ability of the wetland soil to support ecological integrity.

While a number of assessment protocols have been developed for wetlands, assessments based on soil characteristics has proven to be relatively more difficult to estimate and relate to anthropogenic disturbance (Fennessy et al. 2007). Because human disturbance alters wetland soil properties and the biogeochemical cycles through, for example, nutrient enrichment or contaminant accumulation, there is a need for soil-based biogeochemical

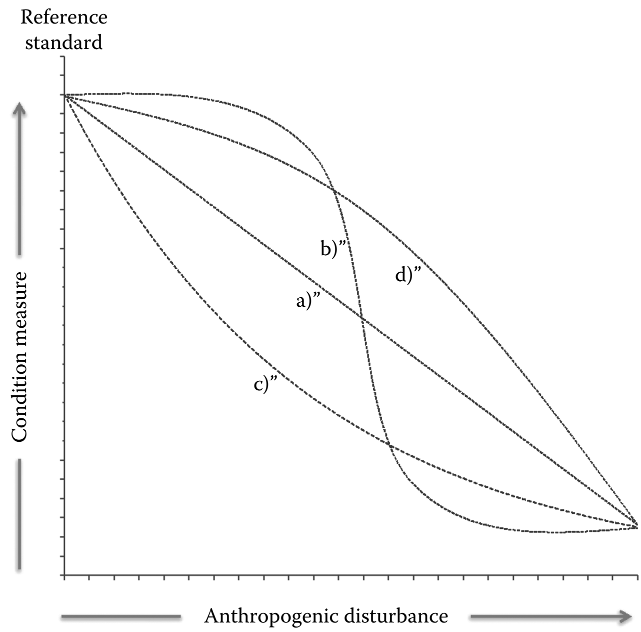


FIGURE 17.2
Concept of reference wetlands along a gradient of anthropogenic disturbance with the reference standard referring to conditions at the least, or minimally impacted sites: (a) linear response of the condition to disturbance; (b) nonlinear response of the condition to disturbance; (c) and (d) potential envelope of the reference wetland condition. (From With kind permission from Springer Science+Business Media: *Mid-Atlantic Freshwater Wetlands: Advances in Wetlands Science, Management, Policy, and Practice*, 2013, 381–420, Brooks, R. and D. Wardrop (Eds.).)

indicators that reflect the structure and function of wetlands (Reddy and DeLaune 2008). Such methods provide an early warning of ecosystem stress, determine the effectiveness of management actions, and track wetland condition for programs charged with management, restoration, and mitigation (Brinson and Malvarez 2002; Junk 2002). Despite the number of methods that have been developed, most of them do not address soil quality or contain indicators based on soil characteristics. In a review, Fennessy et al. (2007) found that among 16 wetland rapid assessment methods evaluated, only two of them used more than one soil parameter as part of the assessment protocol, and most methods had none. Soil indicators that have been used are almost entirely based on qualitative soil morphological features including soil type, visible signs of substrate disturbance (e.g., tire tracks), presence of redoximorphic features, depth of A horizon, or Munsell color. These have limited ability to assess soils compared to more quantitative indicators (Fennessy et al. 2007; Twohig and Stolt 2011). Soil characteristics have promise as the basis for robust indicators; they are sensitive to change; and may reflect disturbance before there are noticeable changes in the biota; thus, they can serve as an early warning of human impacts (Corstanje et al. 2009).

Because there is a need for comprehensive methods that include soils in wetland condition assessments, this chapter presents an approach that has been applied to the development of other assessment methods to identify and evaluate soil indicators. Indicator development generally follows four major steps: (1) the establishment of a standard of ecological integrity (i.e., soil quality) against which to measure, (2) the appropriate measures of soil physical, chemical, or biological properties (i.e., what are the best measures of soil quality), (3) classification of indicators according to the levels of effort and resources required, so that multiple options exist for monitoring and assessment, and (4) the identification of which indicators reflect the general condition, versus those that are relevant to a specific function. Each of these is addressed separately below.

Standards of Soil Quality

The concept of an appropriate standard against which to measure ecological condition requires specification of wetland type and construction of a gradient of disturbance. Wetlands encompass a wide diversity of habitats and soils that vary in their physical, chemical, and biological characteristics (including peat bogs, mineral-rich fens, forested swamps, freshwater and saltwater marshes, and pocosins), which, in turn, leads to differences in the functions or ecosystem services they perform. While by definition wetlands have hydric soils, the characteristics of the soil vary with wetland type, for example, peat-accumulating wetlands tend to have a greater capacity for C storage than do freshwater marshes. Creating classes of similar sites within or across regions reduces variability due to natural differences in soil characteristics, making the effects of human disturbance easier to discern, and the responsiveness of indicators more clear.

Establishing the expectations of condition for a specified wetland type and location is the next step. The reference approach presented by Brinson (1993) denotes a range of wetland conditions that can be correlated with a gradient of anthropogenic disturbance (Figure 17.2). Reference standard refers to the condition at the least-, or minimally, impacted sites and provides the basis for a quantitative description of the best available physical, chemical, and biological properties of the specified wetland type, given the current state of the landscape. The reference concept provides a number of critical elements for describing and understanding the relationship between wetland condition and anthropogenic disturbance (Rheinhardt et al. 2007; Wardrop et al. 2013): it provides the grounding at either end of the condition/disturbance gradient; it defines the nature of the relationship (e.g., linear or nonlinear with thresholds); and it provides guidance as to the expected variability in condition at any value of the disturbance gradient (e.g., differences in the assessed condition of two sites that are subjected to the same level of anthropogenic disturbance). It also articulates three benchmarks that are important in management: minimally disturbed (condition in the absence of significant human disturbance), least disturbed (condition given the best available conditions of the landscape, e.g., wetlands in a developed landscape), and best attainable (the expected condition of least-disturbed sites if best management practices are employed) (Stoddard et al. 2006).

Selection of Indicators and Levels of Assessment Effort

Effective indicators are those attributes that respond predictably and reliably to human disturbance and the related stressor gradients (Karr and Chu 1999). A robust indicator is one that (1) is relatively easy and inexpensive to measure; (2) is sensitive to anthropogenic stress or disturbance; (3) shows a consistent response to stress or disturbance; (4) is ecologically relevant (i.e., it relates to specific ecosystem processes), and (5) has a minimal amount of seasonal and spatial variability (Fennessy et al. 2001; Schloter et al. 2003; Gil-Sotres et al. 2005). Soil-based indicators can be developed with a focus on their physical, chemical, or biological characteristics to provide information on the ecological condition of a site (i.e., has it departed from the reference condition, and if so, by what magnitude), or to estimate one or more of the functions, or ecosystem services, that wetlands provide.

Assessment approaches have been organized into a "three-tier framework" that arranges indicators hierarchically in terms of the degree of effort they require and the spatial scale they address (Wardrop et al. 2007). Soil indicators can be organized in much the same way to describe the sampling and analytical effort they require (Table 17.1; Reddy and DeLaune 2008). They range from basic measurements using readily obtainable data (Level 1), to more in-depth indicators requiring more intense field and laboratory methods (Level 2), to intensive biological and physicochemical measures (Level 3). Level 1 indicators are low-cost, routine measurements that typically have a relatively long response time, making them less sensitive to human disturbance. Levels 2 and 3 are increasingly more complex measures with higher degrees of spatial variability and shorter response times. Each level can be used to validate and inform the others; for example, intensive assessments (Level 3), which are more rigorous and specific in the biogeochemical processes they assess, have been used to validate less detailed Level 1 measures.

Level 1: Rapid Indicators to Assess Condition

Rapid indicators are relatively simple measures of soil characteristics that provide information on the basic physical and chemical and, to a lesser extent, biological characteristics of a site (Table 17.1). These are straightforward to measure and interpret. Some, like soil pH, may not provide much specific information on changes due to human impacts except for specific wetland classes or types of disturbance. Likewise, parameters such as particle size distribution provide information on a soil's physical makeup, but are typically a function of the soil-forming processes and parent material, and are not necessarily responsive to a changing environment.

Soil pH has been used as an effective indicator of hydrological alterations that alter salt marsh soil chemistry. For example, both the initial soil pH and incubation pH (determined by measuring pH following 2 months of moist soil incubation under aerobic conditions, a simple, but not a rapid measure) were found to be significantly lower in tidally restricted (disturbed) salt marshes than in reference sites (Twohig and Stolt 2011). This was attributed to sulfide oxidation in the tidally restricted sites during the longer periods of drying. Salt marshes typically have highly reduced soils with high sulfide concentrations; drying leads to sulfide oxidation and the consumption of hydrogen ions (see Chapter 13). Thus, the incubation soil pH in the surface soils of reference sites was four or less while it was six or higher in the hydrologically altered sites. Initial soil pH values varied similarly (Twohig and Stolt 2011).

An easily measured parameter that has shown usefulness as an indicator is electrical conductivity (μsiemens cm^{-1}), which is a measure of the soluble salts in a soil solution and a common measure of salinity. It has been used, for example, to assess hydrological flux in arid areas that experience drydown, with the associated rapid change in salt concentrations. In the Prairie Potholes, electrical conductivity has been shown to indicate the relative balance of water inflows and outflows that determine salinity. Water is gained and lost in potholes through precipitation, evapotranspiration, and ground water in- and out-flows, the latter of which carry salts into or out of each wetland (see Chapter 3). Net water outflow results in ephemeral to semipermanent ponds that are fresh to brackish, while potholes with net inflow are semipermanent to permanent ponds that are brackish to saline (Sloan 1972).

TABLE 17.1

Potential Soil Biogeochemical Indicators for Assessing Wetland Impacts

	Indicator Type	Description
Level 1: Rapid/Routine Indicators		
Bulk density	Physical	g Soil cm^{-3}
Particle size distribution	Physical	Range and abundance of particle diameters
Soil pH	Chemical	pH units
Electrical conductivity	Chemical	µsiemens cm^{-1}
Total nitrogen	Chemical	mg TN kg^{-1}
Total phosphorus	Chemical	mg TP kg^{-1}
Total inorganic phosphorus (TiP)	Chemical	mg Inorganic P kg^{-1}
Labile inorganic phosphorus (iP)	Chemical	mg Labile-inorganic P kg^{-1}
Extractable nutrients	Chemical	mg L^{-1}
Total carbon	Biological	g Total C kg^{-1}
Organic matter content	Biological	% of organic matter
C:N:P ratios	Biological	Elemental ratio
Level 2: Moderate Intensity Indicators		
Cation exchange capacity (CEC)	Physical	meq 100 g^{-1}, $cmol_c$ kg^{-1}
Soil porewater nutrients	Chemical	mg L^{-1}
Extractable P	Chemical	mg P kg^{-1}
Degree of P saturation	Chemical	% of soil capacity to bind P
Oxalate extractable metals	Chemical	mg kg^{-1}
Extractable NH_4 nitrogen	Chemical	mg NH_4 L^{-1}
MBC (microbial) biomass C	Biological	mg Microbial C kg^{-1}
MBN (microbial) biomass N	Biological	mg Microbial N kg^{-1}
MBP (microbial) biomass) P)	Biological	mg Microbial P kg^{-1}
MBC/TC	Biological	mg Microbial C kg^{-1} mg soil C kg^{-1}
MBN/TN	Biological	mg Microbial N kg^{-1} mg soil N kg^{-1}
MBP/TP	Biological	mg Microbial P kg^{-1} mg soil P kg^{-1}
Potential mineralizable N and P	Biological	mg $kg^{-1}day^{-1}$
Microbial respiration	Biological	mg $g^{-1}h^{-1}$
Level 3: Intensive/Process Indicators		
rRNA sequence analysis	Biological	Phylogenetic relationships
Soil and C accretion rates: Cs-137 profiles	Chemical	g m^{-2} $year^{-1}$
P sorption coefficients	Chemical	mg kg^{-1}
Stable isotopes (N-15)	Chemical	$\delta^{15}N$
β-Glucosidase activity	Biological	mg $g^{-1}h^{-1}$
Phosphatase activity	Biological	mg $g^{-1}h^{-1}$
Dehydrogenase activity	Biological	mg $g^{-1}h^{-1}$
Urease activity	Biological	mg $g^{-1}h^{-1}$
Denitrification	Biological	mg N_2O–N $kg^{-1}h^{-1}$
Methanogenesis	Biological	mg CH_4–C $kg^{-1}h^{-1}$
Cellular fatty acids	Biological	% of dry cell biomass functional groups

Source: Adapted from Reddy, K. R. and R. D. DeLaune. 2008. *Biogeochemistry of Wetlands*. CRC Press, Boca Raton, FL.

Note: The relative level of effort required and the type of each indicator is shown.

Hydrologic conditions that lead to the accumulation of soil C are a defining feature of wetlands (Bridgham et al. 2006; Mitsch and Gosselink 2007). The anaerobic conditions typical of wetlands, together with characteristically high rates of primary production, lead to the accumulation of soil organic matter (SOM) (or organic C). Thus, wetland soils are a major reservoir of organic matter and an important C sink. Compared to agricultural soils that contain an average of 0.5%–2% C (up to 5% C, or 10% OM), wetland soils can accumulate up to 30%–40% C (or nearly 90% OM: Lal et al. 1995). Organic matter also helps mediate, through microbial action, the conversion of nutrients between available and recalcitrant nutrient pools. Thus, soil C concentrations also provide an indicator of nutrient-cycling rates, and are related to nutrient-cycling functions as estimated by hydrogeomorphic models of the functional capacity of a site (Figure 17.3; Berkowitz and White 2013).

Human activities can act to increase or decrease organic matter and its associated soil C, making comparisons with reference wetlands that are important to assess the impact of specific stressors. For example, nutrient enrichment can stimulate primary production, increasing organic matter accretion and C accumulation in wetland soils (Craft and Richardson 1993; Freeland et al. 1999). Disturbances due to hydrologic alterations, such as drainage or plowing, have a large impact on soil C flux, moving C from soil to the atmosphere via oxidation, and lowering soil C concentrations (Bridgham et al. 2006). In general, minimally disturbed wetlands of the same class tend to have a higher proportion of SOM (including greater microbial biomass and enzyme activities, discussed below), which results in lower soil bulk density. Bulk density is a routinely measured physical property that has been included in some wetland assessment methods (van Dam et al. 1998; Innis et al. 2000; Rokosch et al. 2009), in part because it is relatively easy to collect samples in the field and process them in the laboratory. Bulk density tends to increase with

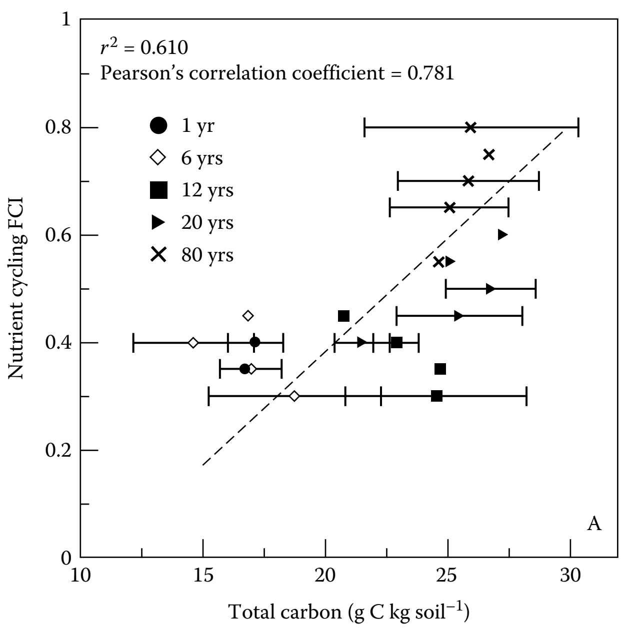


FIGURE 17.3
Comparison of measures of total soil C and the biogeochemical function of nutrient cycling as measured by a rapid assessment method, providing functional capacity index scores. (Error bars are one-standard deviation; from Berkowitz, J. F. and J. R. White. 2013. *Soil Sci. Soc. Am. J.* 77: 1442–1451.)

disturbance, although its response is a function of the type of disturbance. Rokosch et al. (2009) found that bulk density was the lowest in undisturbed forested wetlands, and that values increased with disturbance. Disturbance due to grazing in herbaceous wetlands has also been found to increase bulk density, with associated reductions in infiltration and porosity. In this case, as stocking rates of cattle increased, bulk density increased in tandem (Moreno-Casasola et al. 2012). In salt marshes, bulk density has been used as an effective means to estimate the collapse of marsh peat (Twohig and Stolt 2011). Despite its relatively slow response times, it is a valuable indicator because it shows strong links to human disturbance, it is correlated with other soil nutrients, particularly C, nitrogen (N), phosphorus (P), and sulfur (S), and it is easy to measure.

There has been extensive research on the response of wetlands to eutrophication. Anthropogenic nutrient enrichment from agriculture, and increasingly urban land uses, can lead to chronic ecosystem degradation in which the chemical, physical, and biological processes in soils are altered by nutrient loading, which in turn alters ecosystem functions, species composition, and productivity (Craft and Richardson 1997; Childers et al. 2003; Reddy and DeLaune 2008). This makes reliable indicators of eutrophication particularly important to track the effects of nutrient inputs and act as early warning signals before shifts in vegetation or nutrient cycling occur (Corstanje et al. 2009). The Florida Everglades have been the focus of intense study on the effects of eutrophication, where decades of runoff from agricultural drainage networks have created strong gradients in soil nutrient concentrations. Here, nutrients accumulate in soils near the drainage channels that carry water-borne nutrients in agricultural runoff, and concentrations decline with distance from the channel edge (Figure 17.4; Childers et al. 2003; Wright et al. 2009). High levels of soil P are particularly problematic, because areas that are P enriched show a replacement of the native saw grass (*Cladium jamaicense*) with the invasive cattail (*Typha domingensis*). Despite improvements in surface water quality due to restoration efforts, soil P levels have not substantially declined; concentrations well in excess of 1000 mg P kg^{-1} have been noted for distances of 4 km or more from irrigation canals, with a concomitant expansion of cattail monocultures (Childers et al. 2003). Total P (TP) concentrations of both soil and sediment floc were shown to be responsive to increased P loads; mean TP levels were 140% and 185% higher respectively, in enriched compared to oligotrophic areas, demonstrating that both floc P and soil P are effective indicators of eutrophication (Wright et al. 2009). Increasing soil P in response to excess nutrients in surface waters has been found in many ecosystems (e.g., Craft et al. 2007). In a comparison of soil and vegetation-based indicators, Craft et al. (2007)

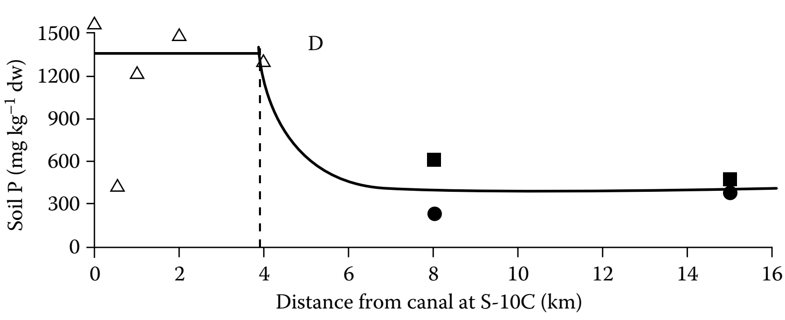


FIGURE 17.4
Soils data from the Everglades Water Conservation Area (WCA) 2A transect indicating soil P levels with distance from a drainage canal and indicating a threshold distance beyond which P levels begin to decline. (From Childers, D. L. et al. 2003. *J. Environ. Qual.* 32: 344–362.)

found that TP was one of the only soil measures to respond predictably to eutrophication across a range of wetlands in the Midwestern United States, showing a positive correlation to surface water phosphate concentrations and the overall trophic status of a site.

The use of soil N concentrations as a basis for biogeochemical indicators is mixed. Total N (TN) has been shown to be an effective indicator of eutrophication, in some cases decreasing as TP increases (Corstanje et al. 2009). However, total Kjeldahl nitrogen (TKN, a measure of organic forms of N) is variable and often does not perform as well as P as an indicator of ecosystem change (Craft et al. 2007). This may be due, in part, to the chemistry of N and the microbial transformations that lead to N loss from the system, for example, through denitrification. At the watershed scale, much of the reactive N that enters a watershed is lost or taken up before it can leave at the outflow (Billy et al. 2010), and much N processing is patchy, occurring within spatially or temporally segregated "hot-spots" or "hot-moments" (McClain et al. 2003). In contrast, N isotopes have shown to be more robust indicators (see below).

It is common to use multiple indicators to assess soil quality because of the complexity of biological and biochemical transformations that occur in soils (Schloter et al. 2003). For instance, assessing P dynamics using combinations of chemical measures such as labile-organic P, and fulvic-, humic-, and labile-inorganic P better reflected nutrient enrichment than did single measures such as TP in Everglades soils. In another example, Richardson and Hussain (2006) found large differences in the physical and chemical characteristics of natural, diked, drained, and reflooded (restored) marsh soils in the Mesopotamian wetlands in Iraq. A multivariate analysis of soil properties using percent C, N, extractable iron and aluminum, soil moisture, pH, salinity, sulfate, and the exchangeable cations calcium and potassium explained 71% of the variability between sites, with clear relationships between the history of disturbance and restoration activities in the marshes (Figure 17.5). Most importantly, the analysis identified the specific combination of soil properties that were most effective at indicating restoration progress by showing which sites were similar to reference wetlands.

Level 2: Moderate Intensity Indicators to Assess Condition

Level 2 indicators require more intensive measurements with more involved field and laboratory methods compared to Level 1 methods (Table 17.1). They are typically of higher cost and greater sensitivity to stressors than are Level 1 indicators, and provide greater insight into the ways human activities affect soil processes and the ecosystem function (Reddy and DeLaune 2008). Examples include extractable and porewater nutrients and measures of microbial biomass and its nutrient content (C, N, and P).

Dissolved substances (nutrients and metals) in soil pore water form the basis for a group of chemical indicators. These reflect the availability of nutrients and contaminants, as well as their flux in a soil system. Typically, nutrients and contaminants move between the soil and water phase as a function of their solubility and concentration. The water content of wetland soils varies from approximately 30% to 50% in mineral soils to up to 95% in organic soils, and most of them is present as free (pore) water (i.e., not held by capillary forces). And unlike whole soil measures that tend to have slower response times, the concentrations of nutrients and metals in pore water, as well as the concentrations derived by soil extraction, respond relatively rapidly to changing environmental conditions. Collecting porewater samples *in situ* can be difficult, requiring special equipment such as porewater equilibrators (USEPA 2008). In contrast, soil extraction methods provide relatively simple estimates of plant-available nutrients, as well as their leachability and

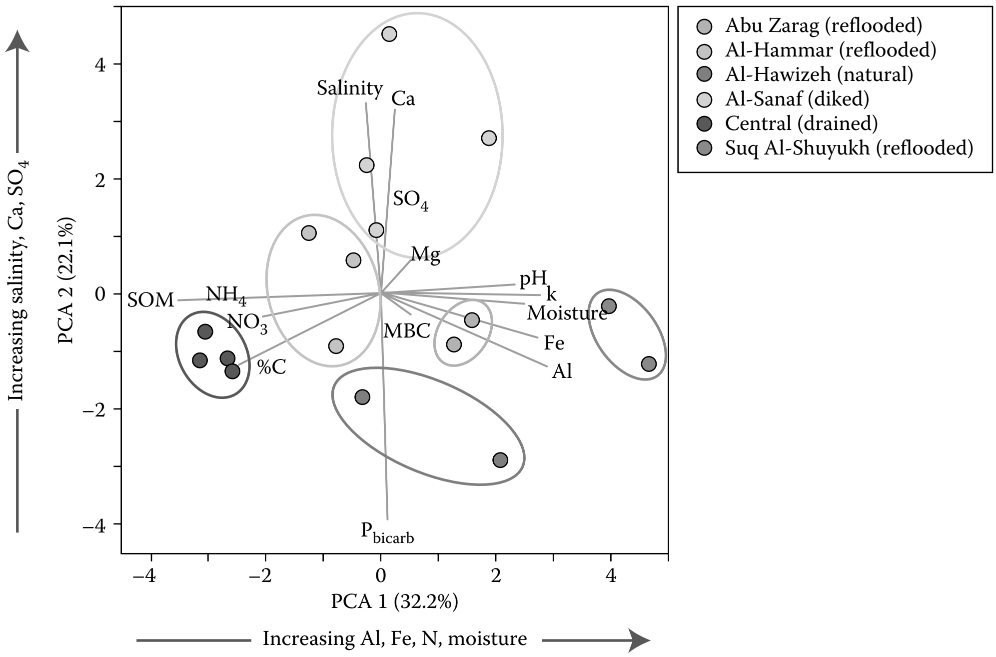


FIGURE 17.5
Principal component analysis (PCA) of Iraqi marsh soils that have been drained, diked, restored, and a natural site. The first two principal component axes account for 52% of the total variance in the data. The soils show strong differences in the characteristics of the marsh sites. (From Richardson, C. J. and N. Hussain. 2006. *Bioscience* 56: 477–489.)

surface runoff potential. These are based on estimates of nutrient release from soils (using water or acid extractions), for example, the release of P or extractable metals into surface waters that are then available for transport downstream (Mukherjee et al. 2009).

In another approach, soil P saturation has been used to predict thresholds above which P will be released to the water column (Richardson 1985). In the southeastern United States, Mukherjee et al. (2009) estimated a threshold value for P saturation, calculated as a function of extractable forms of P, such that

$$\frac{\text{oxalate} - \text{extractable P}}{\text{oxalate extractable Fe} + \text{oxalate extractable Al}} \tag{17.1}$$

Above a value of 0.08, soils became a source of P to the overlying water column. A similar approach has been applied at the ecosystem scale to determine assimilative capacity thresholds for nutrients, beyond which the ecosystem structure and function are altered. In a meta-analysis of data from a large number of wetlands, Richardson and Qian (1999) used areal input and output P mass-loading rates to establish an average P assimilative capacity for wetlands of 1 g P m^{-2} yr^{-1}. Loading rates below this level maintained community structure (e.g., plant community composition) and function (e.g., maintenance of water quality); above this rate, there was a wholesale shift in the plant community composition and altered biogeochemical cycles.

Level 2 indicators have also been developed based on the nutrient content of microbial biomass. Approximately 50% of all microbial biomass is located in the surface 10 cm of the

TABLE 17.2

Microbial Indicators Documented along a Gradient of Nutrient Enrichment in Water Conservation Area 2 of the Florida Everglades (Microbial Biomass C, N, and P; and Potential Mineralizable P, N)

	Nutrient- Impacted Site	Intermediate Site	Unimpacted Site
MBC (g kg^{-1})	7.5 ± 0.7	12 ± 1.3	9 ± 1.0
MBP (mg kg^{-1})	159 ± 9	237 ± 20	73 ± 4
MBN (mg kg^{-1})	1019 ± 91	1709 ± 228	897 ± 99
PMP (mg kg^{-1} d^{-1})	13.5 ± 2.8	5.8 ± 0.2	1.9 ± 0.2
PMN (mg kg^{-1} d^{-1})	42.2 ± 1.2	51.4 ± 2.4	32.3 ± 1.1

Source: Adapted from Corstanje, R. et al. 2007. *Ecol. Indicators* 7: 277–289.
Note: Values are mean ± standard error (n = 36) in the top 10 cm of soil.

soil column, and its C, N, and P content has been shown to be responsive to eutrophication and disturbance (Reddy and DeLaune 2008). For example, microbial biomass C (MBC) is closely linked to C cycling and storage in wetlands, making it a promising indicator. Eutrophication increases microbial biomass carbon (MBC), microbial biomass nitrogen (MBN), and microbial biomass phosphorus (MBP), and empirical data suggest that, among the three, MBC is the most consistent in its sensitivity to disturbance (Qualls and Richardson 2000; Corstanje and Reddy 2006; Rokosch et al. 2009). For instance, in an analysis of Everglades soils, Corstanje et al. (2007) found that MBC, MBN, and MBP increased significantly as nutrient loads increased, with the highest levels at intermediate levels of enrichment. A large number of indicators have been tested for their responsiveness to P enrichment, showing that (Corstanje et al. 2007; USEPA 2008):

- Microbial respiration increased, resulting in a more rapid turnover of organic matter.
- C:P ratios decreased by more than 50% in both detritus and soils.
- The proportion of TP contained in microbial biomass, expressed as the ratio of MBP/TP, declined by 27% in detritus and 50% in soil as P availability increased.
- Higher rates of nutrient turnover in enriched versus oligotrophic sites as measured by potentially mineralizable P and N. P mineralization varied from a high of 13.5 mg kg^{-1} d^{-1} in the most nutrient-rich site, to 5.8 and 1.9 mg kg^{-1} d^{-1} in intermediate and unimpacted sites, respectively (Table 17.2).

Level 3: Intensive Indicators to Assess Condition

Intensive biological indicators (Level 3) are based on detailed biological and chemical information (Table 17.1). Many relate to microbial community composition or the rates of biogeochemical processes, providing information on dynamic soil properties that respond relatively rapidly to changing environmental conditions. Their small size and rapid turnover times make them rapid responders to disturbance resulting from nutrient loading, hydrologic alterations, and contaminant loading (Reddy and DeLaune 2008).

Measures of microbial activity often include measures of extracellular enzyme concentrations, or the products of microbial respiratory pathways including methanogenesis (CH_4 production) or denitrification enzyme activity (DEA; N_2O activity) (USEPA 2008). Extracelluar enzymes are secreted to function outside the cell to aid in the breakdown of high-molecular-weight compounds and the subsequent uptake of mineralized nutrients. They have been shown to be responsive to anthropogenic disturbances that alter wetland soil chemistry in a diversity of wetland types including coastal mangroves (Dinesh et al. 2004), temperate wet forests (Rokosch et al. 2009), and emergent marshes (Wright and Reddy 2001). Enzyme activity, often the rate-limiting step in decomposition, is related to the composition of the microbial community, the distribution of functional groups, and the organic matter content of the soil. For example, the activity of β-glucosidase, which cleaves glucose from larger molecules such as cellulose (cellulose degradation), is a key enzyme involved in the C cycle, and an indicator of decomposition rates. Its activity has been shown to decline with disturbance due to nutrient loading, making it a sensitive indicator of eutrophication (Wright and Reddy 2001). In contrast, β-glucosidase activity did not correlate well with the ecological condition of forested wetlands that spanned a gradient of anthropogenic disturbance; this result was complicated by the inhibition of β-glucosidase activity in sites that experienced longer periods of flooding (Rokosch et al. 2009).

Acid- and alkaline phosphatases are a group of extracellular enzymes used by soil biota to liberate inorganic P from organic compounds. Phosphatase activity (APA) is regulated by the chemical status of the soil; thus, activity is higher in P-limited systems. Human activities that increase the concentration of highly available forms of P, for instance, soluble reactive P, suppress the production of phosphatases, making this enzyme a good indicator of ecosystem P availability. Typically, the highest levels of APA are found in reference sites unenriched in P, and activities decline as nutrient loads increase (Richardson and Qian 1999; Prenger and Reddy 2004; Corstanje and Reddy 2006). In the Everglades, APA increased with distance from nutrient inflows, and activity levels were highest in the interior marsh where P was in shortest supply; APA was inversely related to dissolved inorganic P along this gradient (Figure 17.6; Richardson and Qian 1999; Reddy and DeLaune 2008). APA has also been shown to decrease in the presence of heavy metals (Gil-Sotres et al. 2005), but not to habitat disturbance (Rokosch et al. 2009).

Other extracellular enzymes have been explored for their utility as indicators including urease and dehydrogenase, although much of this research has been done in terrestrial soils. For example, urease catalyzes the hydrolysis of urea and has been primarily used in terrestrial soils as an indicator of soil management practices such as the application of

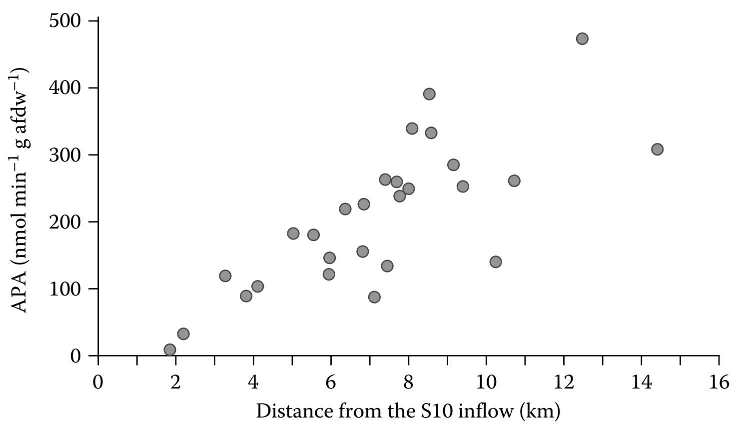


FIGURE 17.6
Relationship between APA and periphyton mats as a function of distance from nutrient inflow in the Everglades. (From Reddy, K. R. and R. D. DeLaune. 2008. *Biogeochemistry of Wetlands.* CRC Press, Boca Raton, FL.)

cattle slurry fertilizers. Dehydrogenase activity is related to redox status and has been used successfully to assess the recovery of degraded soils (Gil-Sotres et al. 2005). In general, the use of extracellular enzymes as indicators of wetland biogeochemistry is an active area of research that will increase the usefulness of these measures as the mechanisms of response are better understood (Knight and Dick 2004).

It can be more challenging to monitor other aspects of microbial community dynamics, for instance, temporal shifts in taxonomic diversity. However, recent developments in molecular techniques have reduced the complexity of the analytical procedures needed to identify and characterize the composition and function of microbial communities, increasing the usefulness of these indicators (Hartmann et al. 2008; Sims et al. 2013). The most common approach to studying microbial diversity is through analysis of 16S rRNA genes that regulate the production of ribosomes in the cell. These are highly conserved with variable regions that can be used to determine taxonomic differences and serve as an indicator of community diversity and function. 16S rRNA has been used in a variety of wetland habitats including coastal ecosystems, alpine meadows, and wetland restoration sites (Hartmann et al. 2008; Sims et al. 2013).

Soil microbial biomass (SMB) and community functional diversity (which identifies groups of microbes that carry out similar functions, regardless of their species identity) have been studied using phospholipid fatty acids (PLFAs) and community-level physiological profiles (CLPPs). PLFA uses the lipid composition of cell membranes both as a measure of biomass and to determine the functional diversity of microbial groups in the soil. The analysis of PLFA provides information on a number of other characteristics of the microbial community such as physiology, taxonomic diversity, and community composition (Rinklebe and Langer 2010). CLPP is a relatively rapid means to characterize microbial communities using whole soil samples by determining the extent to which soil microbes can utilize a diverse set of single C sources. Differences in the use of different C compounds are linked to the functional traits of the community present (Garland 1997; Sims et al. 2013). However, like many approaches that involve culturing samples, the fact that some functional groups cannot be grown *in vitro* limits the proportion of the community that is responsive to the test. Despite this, indicators of soil quality based on microbial functional groups are effective because they are key players in nutrient cycling, they respond rapidly to changes in the soil environment, and they are integrators of the factors that control decomposition and the transformation of nutrients (Corstanje et al. 2009).

Application of Soil Indicators

Table 17.3 presents key soil indicators that have a demonstrated correlation to the level of performance of various wetland functions, or are indicative of specific stressors. The wetland functions listed are those generally considered to be valuable to society (i.e., ecosystem services) and, thus, desired for measurement or approximation. The stressors are the ones that are commonly associated with surrounding agricultural and development land covers that are common across much of the United States. This table can be used as a guide for answering a range of management questions; for example, a goal of maximizing water quality benefits in a given watershed may require an assessment of the nitrate removal capacity of existing wetlands. If merely a general sense of the level of the overall biogeochemical functioning is appropriate, and resources are minimal, then, the Level 1 indicator,

TABLE 17.3

Key Soil Indicators at Different Assessment Levels and Their Links to Various Wetland Functions and Specific Stressors

Function	Level 1	Level 2	Level 3
Hydrologic	Soil pH, electrical conductivity		
Energy dissipation/short-term surface water detention			
Long-term surface water detention			
Biogeochemical	% organic C, bulk density	MBC, MBP, and MBN	16S rRNA, PLFA, and CLPP
Removal of inorganic N	N		^{15}N, DEA
Solute absorption capacity	P	Degree of P saturation; MBC	P sorption coefficients, APA
Retention of inorganic particulates			^{137}Cs
Export of organic C	% organic C, bulk density	MBC	^{137}Cs, β-glucosidase
Habitat	% organic C		
Maintenance of characteristic plant community	% organic C	Degree of P saturation	
Maintenance of characteristic detrital biomass	% organic C	Degree of P saturation; MBC	
Vertebrate community Structure and composition			
Stressors			
Hydrologic alteration	Soil pH		
	Electrical conductivity		
	% organic C		
Nutrient enrichment	% organic C	MBC, MBP, and MBN	APA, β-glucosidase
	Floc P, soil P		
	TN (with mixed results)		
General disturbance	Bulk density	MBC, MBP, and MBN	
	% organic C		

Note: Blank cells indicate that specific indicators have not yet been identified for those functions or stressors.

% of organic C, may be a potentially useful indicator. However, if a site-specific determination of denitrification capacity is desired, and resources are more generous, then, assessment-utilizing nitrogen isotopes ratios (^{15}N) or the denitrification enzyme assay (DEA) will provide a more accurate prediction with less uncertainty.

Indicators of the Biogeochemical Functions Denitrification and C Storage

Denitrification (N processing) and C storage are controlled by a number of factors that operate at a range of spatial and temporal scales, many of which could be the focus of indicator development (Figure 17.7). Denitrification is a microbial process, and as such, it is most directly affected by the factors at the process scale such as the availability of nitrate, dissolved organic carbon (DOC), temperature, pH, and redox potential (Chapter 4; Groffman et al. 2002; Harrison et al. 2011). These process-scale factors are, in turn, affected by the wetland-scale variables of vegetation and hydrology; vegetation affects C availability and temperature while hydrology can affect nitrate loading and redox conditions

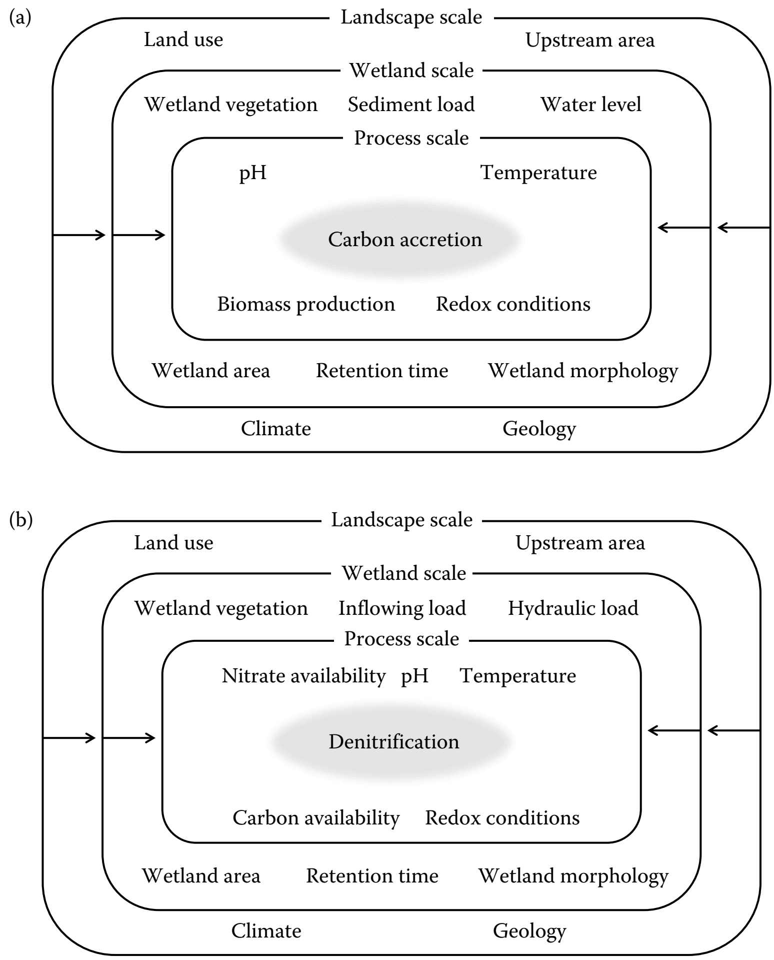


FIGURE 17.7
Factors at various scales affecting the ecosystem function of (a) C accretion and (b) denitrification. (Modified from Trepel, M. and L. Palmeri. 2002. *Ecol. Eng.* 19: 127–140.)

(Adamus and Brandt 1990; Mitch and Gosselink 2007). C sequestration and storage is increasingly recognized as an important function provided by wetlands in the face of rising atmospheric CO_2 concentrations. It is controlled by a hydroperiod, redox potential, primary production, and temperature. Appropriate indicators of these services can be found at all levels depending on the resources available and the goal of the project.

Indicators of C storage include soil C concentrations, soil bulk density (Level 1), and MBC (Level 2), which reflect the ability of a site to sequester and store C. Level 3 indicators of C accretion and storage are often calculated using the presence of radioactive Cesium-137 in the soil profile. Atmospheric deposition of ^{137}Cs, due to aboveground nuclear weapons testing, peaked in 1964; thus, the depth at which ^{137}Cs concentration is the greatest corresponds to the soil surface in 1964 (Craft and Richardson 1993); sediment and organic C accumulation above that depth represents materials deposited since that time. ^{37}Cs is a robust marker because it is strongly absorbed into clay and organic particles, its uptake by vegetation is low, and its diffusion is usually limited (Reddy and DeLaune 2008).

In contrast, denitrification is the primary process by which nitrate is transformed in wetlands, resulting in the removal of N from surface and ground waters. As a microbially mediated process, it is affected by alterations to wetland hydrology, vegetation community structure, and nitrate loading (Figure 17.7b) (Groffman et al. 2002; Mitch and Gosselink 2007). Two functional indicators of denitrification rates have been developed that operate at different temporal scales, the DEA and the proportion of the stable isotope, ^{15}N. DEA is a measure of short-term, localized denitrification rates (i.e., rates at a specific location within a wetland), while ^{15}N integrates denitrification rates over much longer time periods at the ecosystem scale.

The DEA is a short lab incubation that indicates the size and activity of the denitrification enzyme (White and Reddy 1999). Enzyme activity is highly variable and patchy in response to the availability of substrates (organic C, hydrologic inflows that deliver nitrate) creating what are known as "hot-spots and hot-moments" of N processing associated with hydrologic exchanges between surface and subsurface waters, and the residence time of water in those locations (McClain et al. 2003). Rates vary in response to nutrient inflows, for example, DEA ranged from 0.004 to 7.75 mg N_2O–N kg^{-1} h^{-1} in the Florida Everglades, with the highest rates in areas of N enrichment (White and Reddy 1999). DEA provides an effective means to make comparisons of different wetland classes. Because of their connectivity to lotic ecosystems, high C availability, and inflows of nitrate, DEA tends to be higher in riparian and floodplain wetlands compared to depressional sites (Fennessy and Cronk 1997).

The use of the stable isotope ^{15}N as an indicator of denitrification is a relatively new approach based on the ability of isotopes to integrate ecosystem processes over time and space. During denitrification, microbes preferentially take up and use the lighter isotope, ^{14}N, which leaves the soil enriched in the heavier ^{15}N. N isotopic ratios are expressed relative to a standardized reference material that is noted as the ratio, $\delta^{15}N$. Over time, ^{15}N accumulates relative to sites where denitrification rates are lower, increasing the value of $\delta^{15}N$. For example, an early study by Sutherland et al. (1993) found that $\delta^{15}N$ values were higher in depressional wetlands than in upland locations. Other microbial processes affect ^{15}N levels, but denitrification is typically the dominant process affecting a soil's isotopic composition, making $\delta^{15}N$ an effective, semiquantitative indicator of the intensity of denitrification over long, even century-long time periods (Billy et al. 2010).

Indicators to Evaluate Wetland Restoration

Soil quality in restored and created (including mitigation) wetlands has also received considerable attention as the basis for indicators to determine project success (e.g., Bishel-Machung 1996; Shaffer and Ernst 1999; Craft 2001; Fennessy et al. 2008; see Chapter 18). Numerous studies across diverse regions and wetland types have shown that restored and created sites have lower organic C content and nutrient availability. For example, Shaffer and Ernst (1999) measured the SOM concentrations of 95 created palustrine wetlands around Portland, Oregon and found that they contained 60% less SOM than natural sites. Fennessy et al. (2008) found that levels of soil organic C, N, and plant-available P were 21%, 23%, and 61% higher in a series of natural wetlands compared to those created as mitigation for wetland losses. Studies have also documented that SOM and nutrient levels do not necessarily increase over time, as might be expected following restoration (Bishel-Machung 1996; Shaffer and Ernst 1999; Hossler et al. 2011). This is despite the sometimes high rates of primary productivity in created wetlands (Cole et al. 2001). Of concern is that differences in soil nutrients contribute to functional differences through, for example, the ability of soil C to improve soil physical structure that enhances microbial and plant growth (Hossler et al. 2011).

Ecologically sound performance standards based on soil indicators are a critical component of restoration or mitigation programs. Easily measured parameters such as soil organic C and N, and bulk density provide information on the ecological status of restoration projects and are correlated to key ecosystem characteristics such as plant community diversity and nutrient transformations (Hossler et al. 2011). Bulk density is particularly valuable as an integrated measure of soil C, water content, and porosity and is related to many biogeochemical processes such as denitrification, plant and microbial biomass production, and soil C (Figure 17.8), making it an excellent measure of performance. Indicators based on total C and N concentrations have also been developed, for example, mitigation

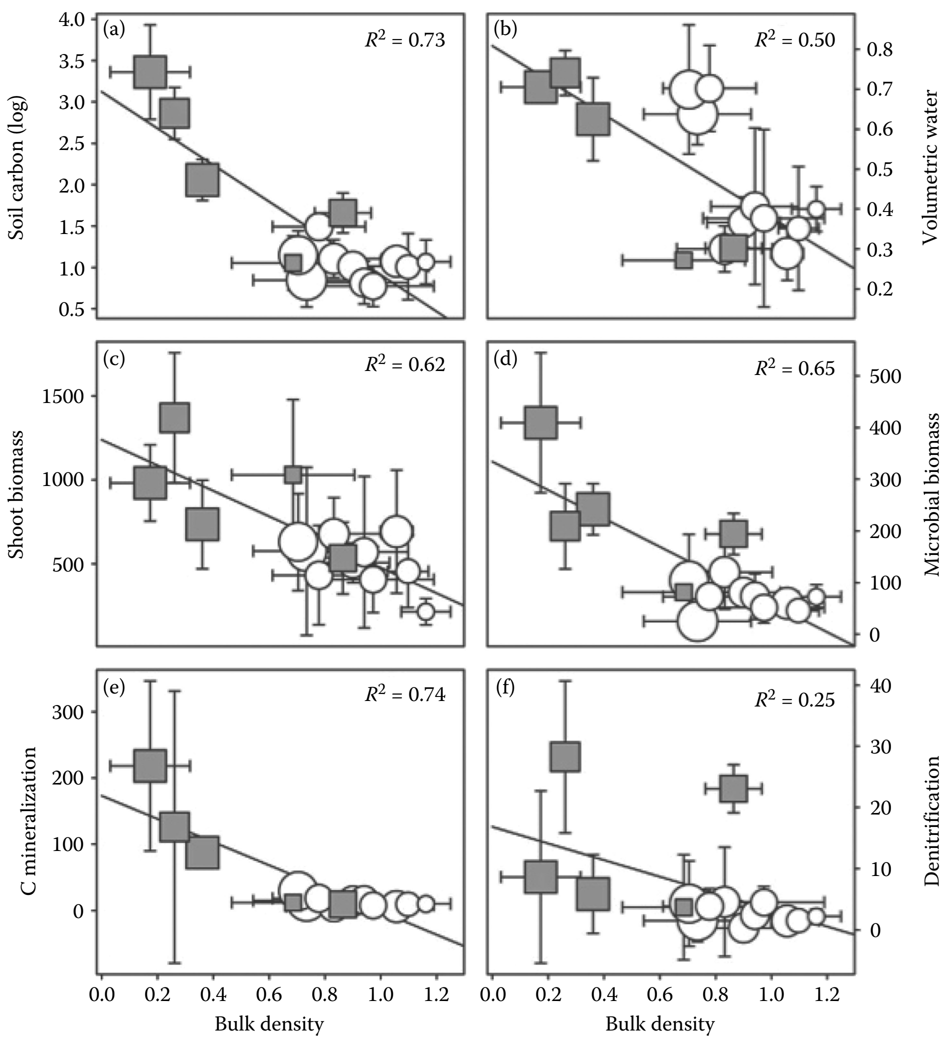


FIGURE 17.8

The correlation of bulk density (g cm^{-3}) with other soil properties and functions, including (a) log-transformed soil C (g kg^{-1}), (b) volumetric water content (m^3 m^{-3}), (c) plant aboveground biomass (g m^{-2}), (d) microbial biomass (nmol g^{-1}), (e) net C mineralization (g kg^{-1} yr^{-1}), and (f) denitrification (mg kg^{-1} yr^{-1}). (Modified from Hossler, K., V. Bouchard, M. S. Fennessy, S. Fry, E. Anemaet, and E. Herbert. 2011. No-net-loss not met for nutrient function: Recommendations for wetland mitigation policies. *Ecospheres* 2: 1–36.)

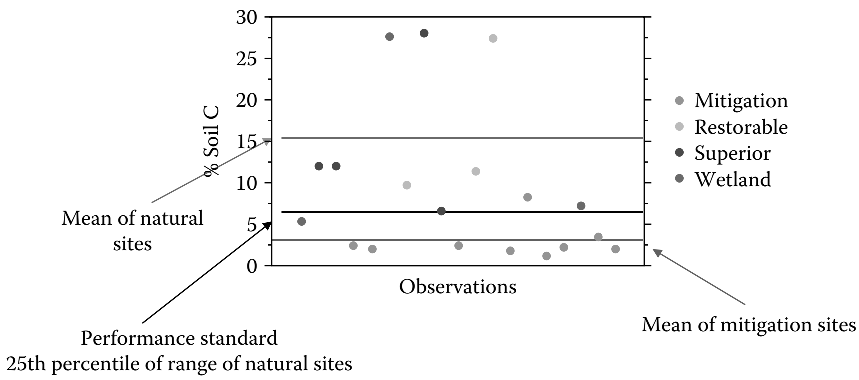


FIGURE 17.9
Translating assessment data to performance standards for soil organic C. In this example, from Ohio wetlands, the mean soil C was 15.1% in natural wetlands and 2.9 in mitigation sites. A performance standard was set at the 25th percentile of the range of natural sites (lines as shown). The ecological status of the natural sites is shown; restorable (low), superior (high), and (moderate) wetland.

standards for the state of Ohio were developed based on the 25th percentile of the range of soil concentrations found in natural wetland soils (6% SOC and 0.5% TN), a relatively conservative target (Figure 17.9; Fennessy et al. 2004). As with bulk density, soil C and N have strong relationships with more complex (and difficult to measure) ecosystem processes.

There is some concern that restored or created sites will show measurable increases toward reference conditions for ecosystem components such as soils. If that is the case, these standards must be met at the time of project establishment, that is, performance standards must be treated as design standards. This will aid in reaching the goals of restoring and maintaining a fully functional and diverse set of wetlands on the landscape.

References

Adamus, P. R. and K. Brandt. 1990. *Impacts on Quality of Inland Wetlands of the United States: A Survey of Indicators, Techniques, and Application of Community-Level Biomonitoring Data*. U.S. Environmental Protection Agency, EPA/600/3-90/073, Environmental Research Laboratory, Corvallis, Oregon.

Berkowitz, J. F. and J. R. White. 2013. Linking wetland functional rapid assessment models with quantitative hydrological and biogeochemical measurements across a restoration chronosequence. *Soil Sci. Soc. Am. J.* 77: 1442–1451.

Billy, C., G. Billen, M. Sebilo, F. Birgand, and J. Tournebize. 2010. Nitrogen isotopic composition of leached nitrate and soil organic matter as an indicator of denitrification in a sloping agricultural plot and adjacent uncultivated riparian buffer strips. *Soil Biol. Biochem.* 42: 108–117.

Bishel-Machung, L., R. P. Brooks, S. S. Yates, and K. L. Hoover. 1996. Soil properties of reference wetlands and wetland creation projects in Pennyslvania. *Wetlands* 16: 532–541.

Bridgham, S. D., J. P. Megonigal, J. K. Keller, N. B. Bliss, and C. Trettin. 2006. The carbon balance of North American wetlands. *Wetlands* 26: 889–916.

Brinson, M. M. 1993. A hydrogeomorphic classification for wetlands. Technical Report WRPDE4, U.S. Army Corps of Engineers, Engineer Waterways Experiment Station, Vicksburg, Mississippi.

Brinson, M. M. and A. Malvarez. 2002. Temperate freshwater wetlands: Types, status and threats. *Environ. Conserv.* 29: 115–133.

Brooks, R. and D. Wardrop (Eds.). 2013. *Mid-Atlantic Freshwater Wetlands: Advances in Wetlands Science, Management, Policy, and Practice*. Springer Publishing, New York, NY, pp. 381–420.

Childers, D. L., R. F. Doren, R. Jones, G. B. Noe, M. Rugge, and L. J. Sciento. 2003. Decadal change in vegetation and soil phosphorus pattern across the Everglades landscape. *J. Environ. Qual.* 32: 344–362.

Cole, C. A., R. P. Brooks, and D. H. Wardrop. 2001. Assessing the relationship between biomass and soil organic matter in created wetlands of central Pennsylvania, USA. *Ecol. Eng.* 7: 423–428.

Corstanje, R., K. M. Portier, and K. R. Reddy. 2009. Discriminant analysis of biogeochemical indicators of nutrient enrichment in a Florida wetland. *Eur. J. Soil Sci.* 60: 974–981.

Corstanje, R. and K. R. Reddy. 2006. Microbial indicators of nutrient enrichment: A mesocosm study. *Soil Sci. Soc. Am. J.* 70: 1652–1661.

Corstanje, R., K. R. Reddy, J. P. Pregner, S. Newman, and A. V. Ogram. 2007. Soil microbial eco-physiological response to nutrient enrichment in a sub-tropical wetland. *Ecol. Indicators* 7: 277–289.

Craft, C., K. Krull, and S. Graham. 2007. Ecological indicators of nutrient enrichment, freshwater wetlands, Midwestern, United States. *Ecol. Indicators* 7: 733–750.

Craft, C. B. 2001. Soil organic carbon, nitrogen, and phosphorus as indicators of recovery in restored Spartina marshes. *Ecol. Restoration* 19: 87–97.

Craft, C. B. and C. J. Richardson. 1993. Peat accretion and phosphorus accumulation along a eutrophication gradient in the northern Everglades. *Biogeochemistry* 22: 133–156.

Craft, C. B. and C. J. Richardson. 1997. Recent and long-term organic soil accretion and nutrient accumulation in the Everglades. *Soil Sci. Soc. Am. J.* 62: 834–843.

Dinesh, R., S. G. Chaudhuri, A. N. Ganeshamurthy, and S. C. Pramanik. 2004. Biochemical properties of soils of undisturbed and disturbed mangrove forests of South Andaman (India). *Wetlands Ecol. Manage.* 12: 309–320.

Fennessy, M. S. and J. K. Cronk. 1997. The effectiveness and restoration potential of riparian ecotones for the management of nonpoint source pollution, particularly nitrate. *Crit. Rev. Environ. Sci. Technol.* 27: 285–317.

Fennessy, M. S., M. Gernes, J. J. Mack, and D. H. Wardrop. 2001. *Methods for Evaluating Wetland Condition: Using Vegetation to Assess Environmental Conditions in Wetlands*. EPA-822-R-02-020. U.S. Environmental Protection Agency, Office of Water, Washington, DC, USA.

Fennessy, M. S., J. J. Mack, A. Rokosch, M. Knapp, and M. Micacchion. 2004. Integrated Wetland Assessment Program. Part 5: Biogeochemical and Hydrological Investigations of Natural and Mitigation Wetlands. Ohio EPA Technical Report WET/2004-5. Ohio Environmental Protection Agency, Wetland Ecology Group, Division of Surface Water, Columbus, Ohio.

Fennessy, S., A. Jacobs, and M. Kentula. 2007. An evaluation of rapid methods for assessing the ecological condition of wetlands. *Wetlands* 27: 504–521.

Fennessy, S., A. Rokosch, and J. Mack. 2008. Decomposition and nutrient cycles in natural and restored wetlands: A functional analysis. *Wetlands* 28: 300–310.

Freeland, J. A., J. L. Richardson, and L. A. Foss. 1999. Soil indicators of agricultural impacts on Northern Prairie Pothole wetlands: Cottonwood Lake Research Area, North Dakota, USA. *Wetlands* 19: 56–64.

Garland, J. L. 1997. Analysis and interpretation of community-level physiological profiles in microbial ecology. *FEMS Microb. Ecol.* 24: L289–L300.

Gil-Sotres, F., C. Trasar-Cepeda, M. C. Leiros, and S. Seoane. 2005. Different approaches to evaluating soil quality using biochemical properties. *Soil Biol. Biogeochem.* 37: 877–887.

Groffman, P. M., N. J. Boulware, W. C. Zipperer, R. V. Pouyat, L. E. Band, and M. F. Colosimo. 2002. Soil nitrogen cycle processes in urban riparian zones. *Environ. Sci. Technol.* 36: 4547–4552.

Harrison, M. D., P. M. Groffman, P. M. Mayer, S. S. Kaushal, and T. A. Newcomer. 2011. Denitrification in alluvial wetlands in an urban landscape. *J. Environ. Qual.* 40: 634–646.

Hartmann, W. H., C. J. Richardson, R. Vilgalys, and G. L. Bruland. 2008. Environmental and anthropogenic controls over bacterial communities in wetland soils. *PNAS* 105: 17842–17847.

Hossler, K., V. Bouchard, M. S. Fennessy, S. Fry, E. Anemaet, and E. Herbert. 2011. No-net-loss not met for nutrient function: Recommendations for wetland mitigation policies. *Ecospheres* 2: 1–36.
Innis, S. A., R. J. Naiman, and S. R. Elliott. 2000. Indicators and assessment for measuring the ecological integrity of semi-aquatic terrestrial environments. *Hydrobiologia* 422/433: 111–131.
Junk, W. J. 2002. Long-term environmental trends and the future of tropical wetlands. *Environ. Conserv.* 29: 414–435.
Karlen, D. L., M. J. Mausbach, J. W. Doran, R. G. Cline, R. F. Harris, and G. E. Schuman. 1997. Soil quality: A concept, definition, and framework for evaluation. *Soil Sci. Soc. Am. J.* 61: 4–10.
Karr, J. R. and E. W. Chu. 1999. *Restoring Life in Running Waters: Better Biological Monitoring*. Island Press, Washington, DC.
Karr, J. R. and D. R. Dudley. 1981. Ecological perspective on water quality goals. *Environ. Manage.* 5: 55–68.
Knight, T. R. and R. P. Dick. 2004. Differentiating microbial and stabilized beta-glucosidase activity relative to soil quality. *Soil Biol. Biochem*. 36: 2089–2096.
Lal, R., J. Kimble, E. Levine, and C. Whitman. 1995. World soils and greenhouse effect: An overview. In R. Lal, J. Kimble, E. Levine, and B. A. Stewart (Eds.). *Soils and Global Change*. CRC Press, Inc., Boca Raton, Florida, MI, pp. 1–7.
McClain, M. E., E. W. Boyer, C. L. Dent, S. E. Gergel, N. B. Grimm, P. M. Groffman, S. C. Hart. et al. 2003. Biogeochemical hot spots and hot moments at the interface of terrestrial and aquatic ecosystems. *Ecosystems* 6: 301–312.
MEA. Millennium Ecosystem Assessment. 2005. *Ecosystems and Human Well-Being: Wetlands and Water Synthesis*. World Resources Institute, Washington, DC.
Mitsch, W. and J. Gosselink. 2007. *Wetlands*, 4th ed. John Wiley & Sons, Hoboken, New Jersey, USA.
Moreno-Casasola, P., H. L. Rosas, and K. R. Medina. 2012. From tropical wetlands to pastures on the coast of the Gulf of Mexico. *Pastos* 42: 185–217.
Moon, J. B. and D. H. Wardrop. 2013. Linking landscapes to wetland condition. In R. Brooks, and D. Wardrop (Eds.). *Mid-Atlantic Freshwater Wetlands: Advances in Wetlands Science, Management, Policy, and Practice*. Springer Publishing, New York, NY, pp. 61–108.
Mukherjee, A., V. D. Nair, M. W. Clark, and K. R. Reddy. 2009. Development of indices to predict phosphorus release from wetland soils. *J. Environ. Qual.* 38: 878–886.
Prenger, J. P. and K. R. Reddy. 2004. Microbial enzyme activities in a freshwater marsh after cessation of nutrient loading. *Soil Sci. Soc. Am. J.* 68: 1796–1804.
Qualls, R. G. and C. J. Richardson. 2000. Phosphorus enrichment affects litter decomposition, immobilization and soil microbial phosphorus in wetland mesocosms. *Soil Sci. Soc. Am. J.* 64: 799–808.
Richardson, C. J. 1985. Mechanisms controlling phosphorus retention capacity in freshwater wetlands. *Science* 228: 1424–1427.
Richardson, C. J. and N. Hussain. 2006. Restoring the Garden of Eden: An ecological assessment of the marshes of Iraq. *Bioscience* 56: 477–489.
Reddy, K. R. and R. D. DeLaune. 2008. *Biogeochemistry of Wetlands*. CRC Press, Boca Raton, FL.
Rheinhardt, R., M. M. Brinson, R. Brooks, M. McKenney-Easterling, J. Rubbo, J. Hite, and B. Armstrong. 2007. Development of a reference-based method for identifying and scoring indicators of condition for coastal plain riparian reaches. *Ecol. Indicators* 7: 339–361.
Richardson, C. J. and S. Qian. 1999. Long-term phosphorus assimilative capacity in freshwater wetlands: A new paradigm for sustaining ecosystem structure and function. *Environ. Sci. Technol.* 33: 1545–1552.
Rinklebe, J. and U. Langer. 2010. Relationship between soil microbial biomass determined by SIR and PLFA analysis in floodplain soils. *J. Soils Sediments* 10: 4–8.
Rokosch, A., V. Bouchard, S. Fennessy, and R. Dick. 2009. The use of soil parameters as indicators of quality in forested depressional wetlands. *Wetlands* 29: 666–677.
Schloter, M., J. C. O'Dilly, and J. C. Munch. 2003. Indicators for evaluating soil quality. *Agric. Ecosyst. Environ.* 98: 255–262.
Shaffer, P. W. and T. L. Ernst. 1999. Distribution of soil organic matter in freshwater emergent/open water wetlands in the Portland, Oregon metropolitan area. *Wetlands* 19: 505–516.

Sims, A., Y. Zhang, S. Gajaraj, P. Brown, and Z. Hu. 2013. Toward the development of microbial indicators for wetland assessment. *Water Res.* 47: 1711–1725.

Sloan, C. E. 1972. *Ground Water Hydrology of Prairie Potholes in North Dakota*. USGS Professional Paper 585-C, Washington, DC.

Stoddard, J. L., D. P. Larsen, C. P. Hawkins, R. K. Johnson, and R. H. Norris. 2006. Setting expectations for the ecological condition of streams: The concept of reference condition. *Ecol. Appl.* 16: 1267–1276.

Sutherland, R. A., C. van Kessel, R. E. Farrell, and D. J. Pennock. 1993. Landscape-scale variations in plant and soil nitrogen-15 natural abundance. *Soil Sci. Soc. Am. J.* 57: 169–178.

Trepel, M. and L. Palmeri. 2002. Quantifying nitrogen retention in surface flow wetlands for environmental planning at the landscape scale. *Ecol. Eng.* 19: 127–140.

Twohig, T. M. and M. H. Stolt. 2011. Soils-based rapid assessment for quantifying changes in salt marsh condition as a result of hydrologic alterations. *Wetlands* 31: 955–963.

U.S. EPA. 2008. *Methods for Evaluating Wetland Condition: Biogeochemical Indicators*. Office of Water, U.S. Environmental Protection Agency, Washington, DC. EPA-822-R-08_022.

Van Dam, R. A., C. Camilleri, and C. M. Finlayson. 1998. The potential of rapid assessment techniques as early warning indicators of wetland degradation: A review. *Environ. Toxicol. Water Qual.* 13: 297–312.

Wardrop, D. H., M. E. Kentula, R. Brooks, M. S. Fennessy, S. Chamberlain, K. Havens, and C. Hershner. 2013. Monitoring and assessment of wetlands: Concepts, case-studies, and lessons learned. In R. Brooks, and D. Wardrop (Eds.). *Mid-Atlantic Freshwater Wetlands: Advances in Wetlands Science, Management, Policy, and Practice*. Springer Publishing, New York, NY, pp. 381–420.

Wardrop, D. H., M. E. Kentula, D. L. Stevens Jr., S. F. Jensen, and R. P. Brooks. 2007. Regional assessments of wetland condition: An example from the Upper Juniata watershed in Pennsylvania, USA. *Wetlands* 27: 416–431.

White, J. R. and K. R. Reddy. 1999. Influence of nitrate and phosphorus loading on denitrifying enzyme activity in Everglades wetland soils. *Soil Sci. Soc. Am. J.* 63: 1945–1954.

Wright, A. L. and K. R. Reddy. 2001. Phosphorus loading effects on extracellular enzyme activity in Everglades Wetland Soils. *SSSAJ* 65: 588–595.

Wright, A. L., K. R. Reddy, and S. Newman. 2009. Micobial indicators of eutrophication in Everglades Wetlands. *SSSAJ* 73: 1597–1603.

Zedler, J. B. 2003. Wetlands at your service: Reducing impacts of agriculture at the watershed scale. *Front. Ecol. Environ.* 1: 65–72.

18

Tidal Wetland Restoration

Ellen R. Herbert, John M. Marton, and Christopher B. Craft

CONTENTS

Introduction

Tidal marshes and mangroves are valued for their ecosystem services, including shoreline stabilization, carbon sequestration, water quality improvement through nutrient retention and transformation, water storage and shoreline protection, and habitat for economically important plants and animals (Boesch and Turner 1984, Peterson and Turner 1994, Chmura et al. 2003, Engle 2011). Approximately one-quarter of the global freshwater and saline tidal wetland area has been lost with current rates of loss estimated at 1%–2% per year (Crooks et al. 2011). Losses are largely attributed to human activities and include sea-level rise and saltwater intrusion (Turner 1997, Day et al. 2000, Craft et al. 2009), aquaculture (Valiela et al. 2001, FAO 2007, Coleman et al. 2008), oil and gas exploration (Turner 1997, Day et al. 2000), changes in sediment delivery to the coastal zone (Turner 1997, Day et al. 2000, Weston 2014), and land-use change (Kelly 2001, Warren et al. 2002, Coleman et al. 2008).

Increasingly, the loss of tidal wetlands is compensated for by the restoration or creation of wetlands. *Restoration* is the reestablishment of conditions that sustain wetland hydrology, vegetation, and soils on sites that historically supported wetlands. Wetland *creation* is the establishment of wetland conditions on upland or dredge materials. In this chapter, we discuss the basic goals and design of tidal wetland restoration and creation with a focus on soil processes. We lead with a brief discussion of the history of wetland restoration and the types of restorations that occur to provide context for the evolving nature of wetland creation and restoration. We then discuss the development of created and restored tidal wetland soils, including the establishment of hydrology, sedimentation, soil processes, and plant–soil feedbacks. We address the observed differences between natural, restored, and created wetland soils and their contribution to ecosystem functions. Finally, we discuss the future of wetland restoration in the context of climate change and increasing population growth.

A Brief History of Tidal Wetland Creation and Restoration

Early tidal marsh creation and restoration efforts were heavily influenced by traditional agronomic practices. They often focused on direct seeding and planting of marsh grasses, particularly *Spartina alterniflora* Loisel (Smooth cordgrass), to stabilize dredge-spoil islands (Broome et al. 1986, 1988, Lewis 2000) or re-vegetate breached diked lands (Harvey et al. 1982). From these studies, tidal range, tidal exchange, substrate texture, and nutrient availability were identified as factors that limit the success of these restorations.

Early wetland creation efforts focused on grading upland areas down to an intertidal elevation and digging artificial channels to mimic natural marshes. These creation projects were often established to mitigate for coastal development activities (Lewis 2000, Kelly 2001). The goal of these projects was to reestablish the hydrology and topographic variation that allow for the development of the reduction–oxidation gradients that support biogeochemical function, complex vegetation structure, and facilitate access by finfish and shellfish. Although the design of created wetlands has become more sophisticated, incorporating microtopography, channel design, sediment additions, and species plantings, time still limits the development of these systems. The young-created tidal wetland sites are often not structurally or functionally equivalent to older, natural wetlands (Zedler and Callaway 1999, Craft et al. 2003).

Because the creation of wetlands requires much more intensive management, more recent efforts have focused on restoration with a heavy emphasis on "self-design." These restorations focused primarily on restoring hydrology to former wetlands with the assumption that, given time, wetland geomorphology, soils, and plant communities will develop (Kusler and Kentula 1990), emphasizing the natural processes of ecological succession. These "self-design" projects rely on establishing appropriate intertidal elevations and hydrology thereby facilitating natural processes of sediment accumulation, pioneer colonization, and the eventual maturation of the site to a fully vegetated marsh platform with developed soils resembling those of proximate natural marshes. Many of these projects focus on the restoration of diked lands used for agriculture by allowing tidal flow to return, ultimately resulting in minimal required management and high rates of success. Tidal flows are returned by breaching levees or removing tidal restrictions, plugging man-made ditches, or removing tide control structures in drained sites (Kusler and Kentula 1990, Pethick 2002). Most mangrove restoration and rehabilitation projects still rely on planting, and over time many restorations resemble forest plantations (Kaly and Jones 1998, Ellison 2000). Rehabilitation of mangrove forests is widely used in developing countries where the local community has a say in how the mangrove resource is managed and utilized (Field 1996).

As the science of wetland restoration has advanced and success rates in small projects have increased, there is greater emphasis on large-scale restorations and reestablishing functions at the landscape scale. The United Kingdom, the Netherlands, and other European nations have experimented with the implementation of a system of "managed realignment" or "depoldering" which relocates dikes as a way to return natural hydrologic regimes and mitigate flooding associated with sea level rise (Anisfeld 2012, Esteves 2014, van Staveren et al. 2014). Projects in the Everglades, Florida (Sklar et al. 2005), coastal Louisiana (DeLaune et al. 2003, Lane et al. 2006), and the Yellow River Delta, China (Cui et al. 2008) seek to restore freshwater flows to reestablish natural flood and salinity regimes as a way of stabilizing or even reversing marsh loss.

We present three case studies in the text that illustrate the progression of wetland restoration and the caveats associated with each approach: dredge spoil wetland creations on the U.S. east coast, the "self-design" breached dike marshes of the San Francisco Bay, and landscape-scale large river diversion restoration in coastal Louisiana.

Case Study: Early Marsh Creation on the Atlantic Coast

Some of the earliest experimental plantings of tidal marsh vegetation occurred along the North Carolina (USA) coast in the 1960s and 1970s. W.W. Woodhouse Jr., E.D. Seneca, and S.W. Broome of North Carolina State University planted smooth cordgrass, *S. alterniflora*, to stabilize dredge material and control shoreline erosion (Woodhouse and Knutson 1982). The experiments, funded by the U.S. Army Corps of Engineers, identified propagation techniques, fertilization requirements, and spacing of plants to establish salt marsh vegetation in the high energy environment of the intertidal zone. Through the years, a number of tidal marshes were planted in the region and elsewhere along the Atlantic and Gulf coasts, many for wetland mitigation (Broome 1988, Streever 2000, Havens et al. 2002, Morgan and Short 2002, Edwards and Proffitt 2003, Fearnley 2008). These marshes persist today and have been the focus of a number of studies documenting the development of the plant community, soils, and trophic structure over time (Cammen 1976, Craft et al. 1988, Moy and Levin 1991, Sacco et al. 1994, Levin et al. 1996, Posey et al. 1997, Alphin and Posey 2000, Craft 2000, Craft and Sacco 2003, Craft et al. 2003).

Development of soil properties characteristic of wetlands, especially organic matter enrichment, begins as soon as hydrology is established and vegetation covers the site. An organic-rich surface layer begins to develop and, within 20 years, a 10-cm thick layer is evident (Figure 18.1). Accumulation of organic carbon (C) and nitrogen (N) proceeds faster in surface than in subsurface layers and the increase in both is linear over time. Accumulating organic matter supports the heterotrophic food web, microbial communities, and benthic infauna that form the base of the tidal marsh food web (Craft and Sacco 2003, Craft et al. 2003). Microbial processes such as decomposition (CO_2 evolution), methanogenesis and denitrification are positively and linearly related to soil organic C (Figure 18.2). The density and diversity of benthic infauna increase rapidly with soil organic C but reach an asymptote with about 2% soil organic C (Figure 18.3). The presence of adequate soil organic matter (SOM) is essential to support the tidal marsh food web and thresholds of 1000 g C m^{-2} and 100 g N m^{-2} are needed to achieve this.

Tidal Wetland Restoration—The Role of Hydrology, Soils, and Vegetation

Restoring Hydrology

Restoration of hydrology focuses on two main goals: establishing appropriate intertidal elevations for the establishment of wetland vegetation, and facilitating tidal exchange of

FIGURE 18.1
(**See color insert.**) Development of an organic-rich surface layer 20 years following marsh creation. (Photo provided by C.B. Craft.)

sediment, organisms, and other materials. Hydrology is typically established by grading and the creation of tidal creeks in created sites or by removing water control structures (ditches, dikes, and levees) and is critical for setting the pace of ecosystem recovery following restoration. Hydrology in tidal systems controls hydroperiod, accretion and sedimentation, and water and material exchange.

Hydroperiod—Tidal inundation determines the depth, duration, and frequency of flooding that structures oxidation-reduction (redox) gradients and sedimentation processes. Natural marshes develop on a relatively flat plain between mean high water (MHW) and mean higher high water (Figure 18.4; Redfield 1972, Zedler and Callaway 1999, Morris et al. 2002). Restoration projects established too low in the tidal frame may convert into open water or mudflat whereas those too high in the tidal frame often fail to support wetland vegetation or important redox reactions. In subsided systems, simple levee breaching without restoring soil surface elevation results in the conversion of sites to open water because vegetation cannot establish below a certain depth, usually mean lower high water (Figure 18.4; Orr et al. 2003, Miller et al. 2008). It is preferable to fill or grade sites to elevations below those observed in natural marshes to allow for natural processes of sedimentation and organic matter accumulation to develop.

Dredge spoil may be added to subsided sites to increase marsh surface elevation and provide secondary benefits such as mineral and nutrient subsidies (Turner et al. 1994, Ford et al. 1999, Warren et al. 2002). The application of dredge spoil is used both to increase the elevation of diked marshes prior to restoration (Marcus 2000) as well as to subsidize existing marshes. The application of sediment to marshes in Louisiana has increased elevation and facilitated *S. alterniflora* regrowth (Ford et al. 1999, Schrift et al. 2008). Mendelssohn and Kuhn (2003) found that sediment applications increased iron (Fe) and manganese (Mg) concentrations and decreased the accumulation of toxic sulfide and ammonium, ultimately resulting in more favorable conditions for plant development and growth.

Alternatively, some projects utilize culverts, pumps, or tide gates to introduce muted tidal regimes to subsided land. These controlled, reduced tides allow natural accretion

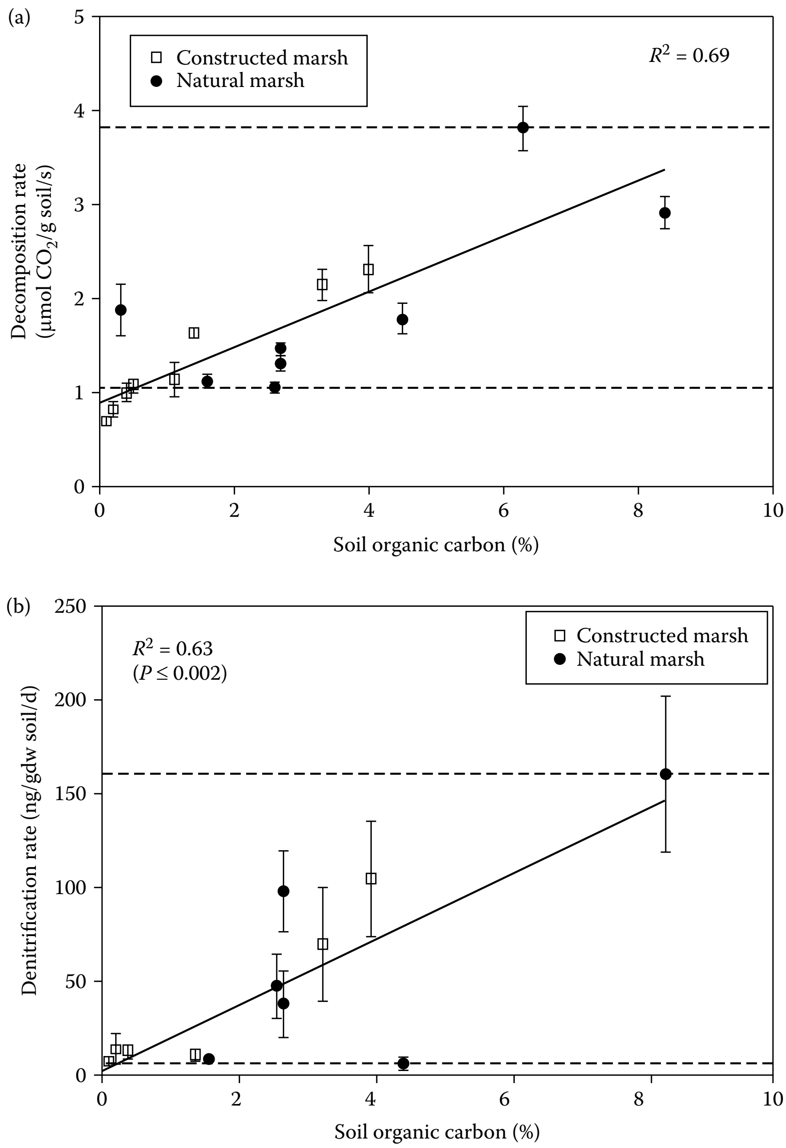


FIGURE 18.2
Relationship between soil organic C and (a) decomposition and (b) denitrification in constructed and natural tidal marshes of North Carolina.

processes, both mineral (Vandenbruwaene et al. 2011) and organic (Miller et al. 2008) to build elevation prior to full reintroduction of tides. Generally, these reduced tides are shallower and longer than the natural tides (Vandenbruwaene et al. 2011), resulting in higher rates of sediment deposition and lower rates of decomposition (discussed below) until desired elevations are reached and then natural tidal regimes are introduced.

Local topography and slope are also important in creating adequate marsh area and heterogeneous surfaces. Without the proper slope, estimated to between 1% and 3% for southeast coast marshes (Broome et al. 1988), there is a less chance of successful colonization of

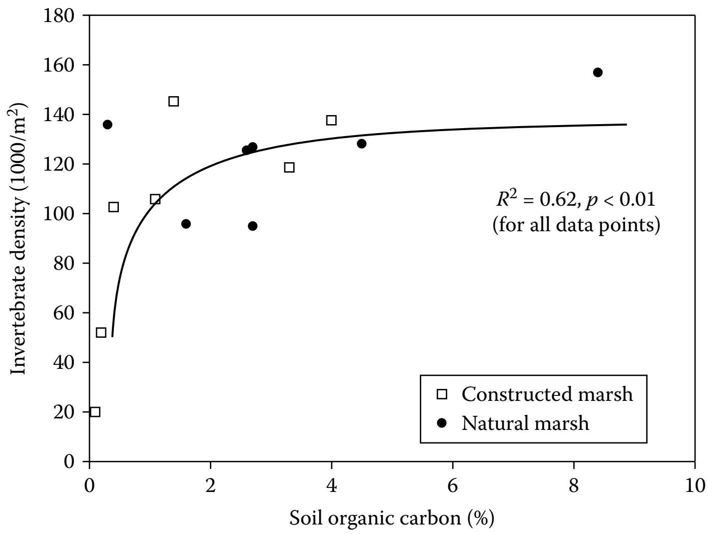


FIGURE 18.3
Relationship between soil organic C and benthic infauna density in constructed and natural tidal marshes of North Carolina.

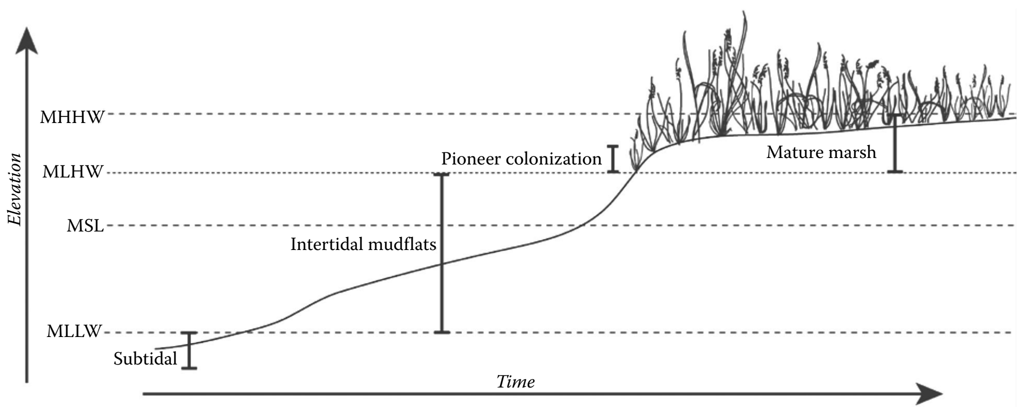


FIGURE 18.4
Theoretical model of tidal wetland development. (Modified from Williams, P. B. and M. K. Orr. 2002. *Restoration Ecology* 3: 527–542.)

marsh vegetation. If the slope is too steep, only a narrow zone will be available for colonization. If the slope is too gentle the site will not drain readily and will become water-logged.

Accretion and Sedimentation—Tidal wetlands are depositional environments where the accumulation of mineral sediment and autochthonous organic matter drives accretion and soil development. Elevation relative to the tidal frame, hydroperiod, and suspended sediment concentration control the rate at which mineral sediment is deposited, while plant growth and decomposition control the accumulation of organic matter *in situ* (Morris et al. 2002). Mineral sedimentation often returns quickly once elevation has been reestablished (Reed 2002). Sedimentation is greater in areas that flood more frequently, such that initial rates of sedimentation in subsided wetlands are generally higher than those in adjacent natural marshes.

Once vegetation has become established, accretion proceeds more rapidly as vegetation stabilizes soils to prevent erosion, baffles water flow, decreases water velocity and increases sedimentation, and adds organic matter to soils (Craft et al. 2003). As marshes approach the elevation of surrounding mature marshes, accretion rates slow, approximating those of nearby natural marshes (Eertman et al. 2002, Morgan and Short 2002, Craft et al. 2003).

Water and Material Exchange—Natural marshes are dissected by complex networks of sinuous, branched channels that facilitate the efficient movement of water, sediment, nekton, and other materials on and off the marsh surface. These tidal channels are difficult to re-establish in marshes and attempts to create them have largely resulted in low density, linear channels which do not function like their natural counterparts. As discussed above, grading sites below the minimum depth for vegetation colonization promotes natural accretion processes and has been found to promote the development of natural drainage channels (Callaway et al. 2011). Newly forming "young natural marshes" at low elevation have a poorly developed drainage network. However, over time, this network develops to resemble those found in natural marshes facilitating elevation increases (Osgood and Zieman 1993). Dampened hydrology and elevations too high in the tidal frame impede the development of proper drainage networks (Eertman et al. 2002).

It is often impractical to completely remove dikes, levees, and other impediments to water flow. Breaches that are too small constrict flow, dampening the tidal signal, and restricting sediment exchange. Although this may be desirable in projects where land has subsided, this can restrict vegetation establishment and stunt the evolution to a fully vegetated marsh plain (Vandenbruwaene et al. 2011).

Breaching levees to restore hydrology brings dramatic changes in tidal inundation and, oftentimes, salinity. Diked marshes often are fresher and more anoxic than adjacent tidal waters as freshwater collects in these impoundments and tidal flushing is restricted or inhibited (Portnoy 1999). Breaching impoundments often increases salinity and alters the redox state of the marsh (Portnoy 1999, Anisfeld 2012). An unintended consequence of the rapid reintroduction of seawater is an increase in sulfate reduction which accelerates organic matter mineralization and marsh subsidence and impedes the restoration effort (Portnoy 1999). Evaluating multiple salt marshes in Connecticut that were restored by breaching dikes, Warren et al. (2002) found that salinity was the primary factor driving *Phragmites australis* (common reed) replacement by *S. alterniflora*. Sites that received less tidal flushing, and were thus fresher, had greater abundances of *Phragmites* and took longer (up to two decades) to exhibit a functional equivalence to reference salt marshes. *Spartina* was also slower to recover in areas with higher elevation that received less frequent and intense tidal flooding.

Re-Establishing Vegetation

Restoration of tidal wetland-dependent functions requires re-establishing hydrology but the development of wetland soils is also tightly coupled to the development of the vegetative community. Vegetation is essential for the physical stabilization of marsh soils and the development of important ecosystem processes including carbon sequestration, denitrification, and support of the detrital food web. Vegetation baffles incoming water, reducing water velocity and increasing sedimentation (Morris et al. 2002). Tidal wetlands are highly productive ecosystems and due to low rates of decomposition, a large portion of organic matter from wetland vegetation is buried in wetland soils. This organic matter builds elevation, reduces bulk density, allowing for greater exchange of materials in the soil profile, sequesters C, N, and P, and provides a source of energy for microbial metabolism and soil infauna.

While early wetland creation efforts relied on plantings to establish vegetative communities (Broome et al. 1988, Craft et al. 1999, Zedler and Callaway 1999, Zedler et al. 2001), in wetland restorations, the vegetation community often reestablishes on its own (Baldwin and Derico 1999, Leck 2003). Seeding and transplanting sprigs, seedlings, and plugs on bare substrates can lead to biomass and areal coverage equivalent to natural marshes within three growing seasons (Broome et al. 1988, Craft et al. 2002). The density of plantings may depend on many factors, including tidal energy and desired time to site coverage. Generally, more dense plantings result in more rapid achievement of full site coverage but require greater initial investment. Planting density may also be affected by physical factors such as tide range and wave energy. In the case of mangroves, mature propagules are planted at densities of less than 5 per m^2 in protected sites. In more exposed sites they are planted at 15 per m^2 (Imbert et al. 2000) or greater densities to compensate for greater exposure and wave energy (Saenger 2002). A number of marsh and mangrove species are planted (Table 18.1) and the species selected will vary depending on geographic region, climate and surface water salinity.

On sites where natural colonization occurs, establishment of good coverage of vegetation is slower, but generally develops within 10 years. On dredge-spoil created wetlands, the macrophyte community has been shown to reach biomass and composition equivalent to natural sites in 5–10 years (Craft et al. 1999, 2003), while in Connecticut, coverage of salt marsh vegetation increased from 0.3% to 8.6% each year following reintroduction of tidal inundation (Warren et al. 2002). As the vegetation community shifted from *Phragmites* to *S. alterniflora* and *Distichlis spicata* L. (salt grass), the abundance and diversity of fish, invertebrates, and birds also increased. In restored or created wetlands relying on natural colonization, low densities of seeds in the seed bank from dredged or excavated soils are compensated for by rapid colonization of ruderal natives and exotic species. Aboveground biomass development can be rapid with species richness exceeding that of natural marshes within a single growing season (Leck 2003, Baldwin 2004). Even in planted marshes, physical factors such as elevation, climate, and tide range may drive marshes toward different plant successional trajectories than are targeted for restoration (Edwards and Proffitt 2003, Osland et al. 2012.)

Nitrogen and sometimes P are needed to establish good coverage of vegetation (Broome et al. 1988) and sometimes maintain it (Boyer and Zedler 1998). This is especially true with sandy soils that are low in organic matter content and have a lower cation exchange capacity (Broome et al. 1988, Zedler and Callaway 2000). A minimum of 100 g N m^{-2} is necessary

TABLE 18.1

Common Marsh and Mangrove Species Planted to Restore Tidal Wetlands

Marsh	Mangrove
Spartina alterniflora	*Rhizophora* spp.
S. cynosuroides	*Avicennia* spp.
S. foliosa (U.S. Pacific coast)	*Sonneratia caseolaris*
S. anglica (Europe)	*Bruguiera gymnorhiza*
S. townsendii (Europe)	*Kandelia candel*
S. patens	*Laguncularia racemosa* (Caribbean)
Juncus roemerianus	*Conocarpus erectus* (Caribbean)
J. gerardii	
Halimioni (Atriplex) portulacoides (Europe)	
Puccinellia maritima (Europe)	

to support a productive, self-sustaining plant community in salt marshes of the southeastern United States (Craft et al. 2003) which takes between 5 and 15 years depending on the rate of N accumulation in soil (discussed further below).

Case Study: Restoring Hydrology to Impoundments on the Pacific Coast

The San Francisco Bay estuary has experienced extensive marsh loss (approximately 90%), leading to widespread efforts to restore tidal wetlands over the last few decades (Williams and Orr 2002). The majority of recent restorations in the Bay have focused on breached levee salt marshes, relying largely on "self-design" principles to reestablish vegetation, sedimentation patterns, and tidal creeks. Over 40 years of restoration has led to several important conclusions regarding breached levee restorations, including the importance of initial elevation, breach size, adequate sediment supply, and size of the breached area (Williams and Orr 2002).

The ongoing South Bay Salt Ponds restoration will restore 15,100 acres of industrial salt production ponds to a combination of open water and marsh habitat, relying largely on natural sedimentation. Two ponds breached early in the restoration, ponds A6 and A21, show high accretion rates (Callaway et al. 2009), and are rapidly progressing toward vegetated marsh plains (Figure 18.5). Given the high rates of sedimentation required to reestablish marshes in the South Bay, there is concern that with continued restoration of the remaining pond area, sediment supply will begin to limit accretion (Callaway et al. 2009, 2012, Brew and Williams, 2010). Brew and Williams (2010) used a hybrid morphologic model-sediment budget analysis and concluded that while marsh restorations will not become sediment limited, there will be a significant decrease in proximal mudflats.

The Bay also contains a substantial number of restoration sites that are considered too low to breach without prior intervention. In some cases, draining, tilling, or other activities have resulted in land subsidence of up to 6 m. The placement of fill is necessary to bring site elevations up before dikes are breached (Marcus 2000, Williams and Orr 2002).

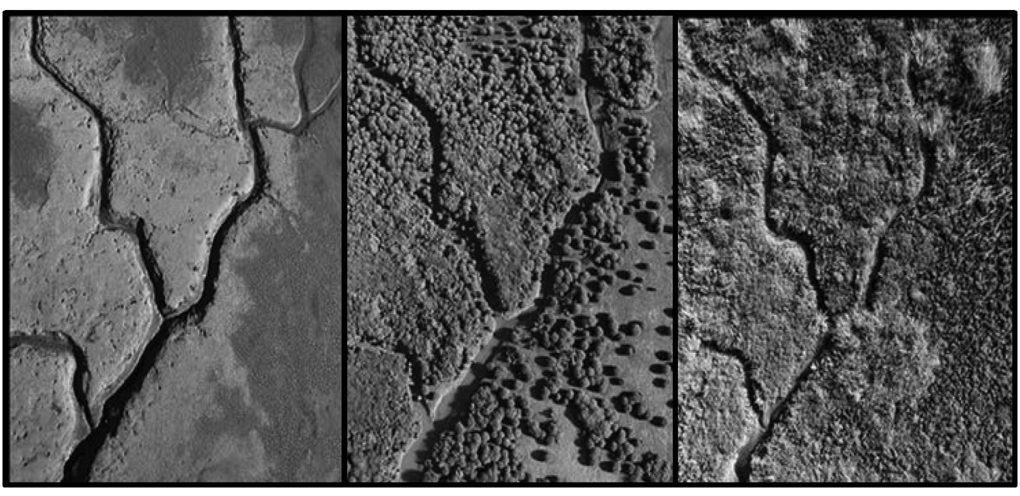


FIGURE 18.5
(**See color insert.**) Aerial photographs of the same creek network at Pond A21 in April 2008 (2 years post breach), September 2009 (3.5 years post breach), and June 2011 (5 years post breach) illlustrating the rapid sedimentation and natural re-vegeatation of the marsh. (Photo by Chris Benton.)

The Sonoma Baylands project utilized 2 million m^3 of dredge material to build elevation slightly below MHW (Marcus 2000). For freshwater peat marshes in the Bay, experiments designed to build peat before dike breaching by first creating nontidal wetlands with controlled hydrology are ongoing and show that an average of 4 cm yr^{-1} of elevation gain can be achieved through organic accumulation (Miller et al. 2008).

Soil Development

Controlling Factors

Although the establishment of appropriate hydrology and vegetation can accelerate soil development, the soils of restored and created wetlands are still limited by time. Following establishment of hydrology and vegetative cover in restored and created wetlands, soil often quickly develops the predominantly anaerobic conditions and redoxomorphic features observed in natural wetlands. Previously reduced soils, such as formerly impounded wetlands and highly reduced dredge spoils, often develop these characteristics more rapidly than those on upland soils. Overall, soil development proceeds faster on degraded sites restored by re-introducing tidal inundation where hydric (wetland) soils still exist, whereas marshes created from dredged material are slower to develop. Marshes constructed on graded terrestrial soils are the slowest to develop since they lack the defining characteristics of topsoil, including sufficiently high organic matter and N content and sufficiently low bulk density to facilitate plant colonization. In created wetlands, upland soils are graded to intertidal wetland elevations (Callaway et al. 2011) often exposing the poor, coarsely textured B and C horizons of upland soils (Craft et al. 1991). Furthermore, soils which have undergone heavy manipulation during site construction can be highly compacted (Bantilan-Smith et al. 2009). Like soils derived from dredge spoil, these soils are low in organic matter and N, and have higher bulk density and lower porosity.

The accumulation of organic matter contributes to the physical integrity of marsh soils by increasing porosity, air and water movement, and supports heterotrophic food webs. Soil organic matter and N accumulate over time after restoration but it is slow (Craft et al. 1988, Simenstad and Thom 1996, Zedler and Callaway 1999, Craft et al. 2002, Zedler et al. 2001). Organic matter accumulation and incorporation into the soil profile is confined to the surface layers in wetlands created or restored on new mineral substrates. Coastal wetlands have largely developed over the last 5000 years during a time of slow sea level rise. These wetlands develop vertically, building soil upward as water levels increase through the addition of root biomass, senesced tissue, and mineral sediment accumulation (Redfield 1972, Morris et al. 2002). Over time the baffling effect of marsh vegetation traps smaller suspended particles, resulting in observed increases in silt- and clay-sized particles (Redfield 1972, Craft et al. 2003). The addition of organic matter is primarily responsible for decreases in bulk density and the accumulation of C and N.

The accumulation of macro-organic matter (MOM), the mixture of living and dead plant material in the soil profile, develops slightly slower than aboveground biomass, requiring 10–15 years to achieve equivalence with natural marshes (Craft et al. 2003). The decomposition of this MOM is largely responsible for organic C, N, and P accumulation in soils and both constructed and natural wetlands show similar proportions of belowground biomass burial (9% and 6%, respectively). Plant species contribute differentially to the development of soil and ecosystem processes. In southern California in an experiment investigating the benefits of species-rich plantings, strong performers such as *Spartina* and *Salicornia* contributed to higher productivity and greater N accumulation (Zedler et al. 2001, Callaway et al. 2003). For example, *Salicornia virginica* planted alone was able to accumulate as much belowground biomass and N as mixed plantings containing three other species (Callaway et al. 2003).

The high pH (8.2) of seawater and accumulation and decomposition of organic matter, along with the unique set of redox reactions that occur in anaerobic wetland soils, drive soils toward a neutral pH. Bantilan-Smith et al. (2009) observed that restored wetlands had pH values much more similar to natural wetlands than created wetlands in the same area. While C accumulation in created and restored tidal marshes is comparable to accumulation rates in natural marshes (Craft et al. 2003), N accumulation often is greater in restoration marshes. Natural as well as constructed marshes are widely observed to be N limited (Valiela and Teal 1974, Ket et al. 2011) and high rates of N retention in young marshes reflect a greater demand for N in these systems (Craft 2001, Craft et al. 2003) with 95% of N being organic rather than inorganic (Craft et al. 1991). Patterns of P accumulation in created and restored wetlands are highly variable. While a proportion of P is associated with organic matter, P inputs in mineral forms often dominate P accumulation (Craft 1997). While process rates, particularly accumulation rates, may meet or exceed rates in natural marshes, the size of C and N pools are often smaller in created and restored wetlands (Cammen 1976, Langis et al. 1991, Simenstad and Thom 1996, Streever 2000, Havens et al. 2002, Morgan and Short 2002, Craft et al. 1993, Edwards and Proffitt 2003, Fearnley 2008). Soil P, on the other hand, usually is present in adequate supply from deposition of P-rich sediment and high dissolved phosphate in seawater (Craft et al. 2003).

As a result of these upward pedogenic processes in young created, restored, and naturally occurring wetlands, the accumulation of organic matter and nutrients and the change in particle size occurs in the uppermost soil layers (Craft et al. 1992, Osgood and Zieman 1993, Krull and Craft 2009), while the lower layers remain largely unmodified. Soils of restored and created tidal wetlands often have higher bulk density and contain less SOM and N than natural marshes, especially in deeper horizons (Craft et al. 1988, Langis et al. 1991, Zedler and Callaway 1999, Craft et al. 2002, Craft et al. 2003, Edwards and Poffitt 2003, Bantilan-Smith et al. 2009).

Trajectory of Development

Studies of tidal marsh and mangrove restoration projects revealed marked differences in soil properties with natural systems (Table 18.2). Typically, bulk density is greater and organic matter and N is lower in constructed marsh and mangrove systems and these differences persist even after 30 years (Table 18.2). Soil properties of a 42-year-old restored tidal marsh in Georgia, though, were similar to natural marshes in the area. Development of wetland soil properties proceeds faster in surface than subsurface horizons (Table 18.2) as plant roots, concentrated near the soil surface, grow and senesce and organic matter is incorporated.

Soil organic matter and N increased after 11 years in a constructed marsh in San Diego Bay, California, though were still lower than the natural reference marshes (Zedler and Callaway 1999). On the Atlantic coast (North Carolina), even after 28 years, created and restored marsh soils (0–30 cm depth) contained less organic C and N than natural tidal marshes in the region though both elements exhibited increasing trajectories with time (Craft et al. 2003).

Accumulation of SOM and fine-texture particles increases after wetlands are planted and become established (Craft et al. 2003). Ren et al. (2008) compared SOM in 4- to

TABLE 18.2

Soil Bulk Density, Organic Matter, and Total N Created, Restored and Natural Marsh Soils

	Age (Years)	Bulk Density (g cm^{-3})		Organic Matter (%)		Nitrogen (%)	
		Restored	Natural	Restored	Natural	Restored	Natural
Salt Marsh							
North Carolina (USA)[a]	1	–	–	0.03–0.12	0.2–1.3	–	–
North Carolina (0–10 cm)[b]	1–28	0.49–1.43	0.40–1.25	0.3–8	0.6–17	0.02–0.27	0.02–0.48
(10–30 cm)		0.98–1.50	0.35–1.32	0.2–1.2	0.6–21	< 0.01–0.05	0.02–0.53
North Carolina[c]	5–15	1.21–1.35	0.13	0.6–1.2	64	0.05–0.09	1.69
Georgia[d]	42	–	–	4.7	3.8	0.41	0.34
Virginia (0–2 cm)[e]	12	–	–	0.02	0.02–0.04	–	–
(14–16 cm)		–	–	0.01	0.03		
Maine and New Hampshire[f]	1–14	–	–	1–15	5–33	–	–
Louisiana[g]	5–18	–	–	2–19	21	–	–
Louisiana[h]	3–19	1.28	0.66	0.7	9.0	0	0.21
Texas[i]	1	–	–	0.2–1.3	0.4–4.8	–	–
Texas[j]	1–2	–	–	5–13	14	0.11–0.28	0.34–0.42
California[k]	4	–	–	0.2–2.2	4–5	0.01	0.02
Washington[l]	7	–	–	2.5	3.5–8.8	–	–
Mangrove:							
Florida (USA)[m]	4–15	0.66–0.70	0.17–0.23	10–12	32–68	–	–
Florida (USA) (0–10 cm)[n]	2–20	1.05	0.28	6	31	0.16	0.80
(10–30 cm)	1.40	0.47	2	19	0.02	0.44	
Qatar[o]	10	–	–	2.4	3.0	–	–
South China[p]	4–10	–	–	1–2	4	0.05–0.14	0.15

[a] Cammen (1976), dredge spoil marsh.
[b] C. Craft, unpublished data, organic matter expressed as organic C*2.
[c] Craft et al. (2002), excavated marsh. Organic matter expressed as organic C*2.
[d] Craft (2001), restored marsh.
[e] Havens et al. (2002), excavated marsh, organic matter expressed as organic C*2 in g cm^{-3}.
[f] Morgan and Short (2002), constructed marshes.
[g] Fearnley (2008), dredge spoil marshes.
[h] Edwards and Proffitt (2003), dredge spoil marshes.
[i] Lindau and Hossner (1981), dredge spoil marsh.
[j] Armitage et al. (2014), dredge spoil marshes.
[k] Langis et al. (1991), constructed marsh.
[l] Simenstad and Thom (1996), excavated marsh.
[m] McKee and Faulkner (2000), planted forests.
[n] Osland et al. (2012), restored forests.
[o] Al-Khayat and Jones (1999), planted forest.
[p] Ren et al. (2008), planted forests.

10-year-old stands planted with nonnative *Sonneratia apetala* (Buch-Ham) in South China. Soil organic matter increased with stand age, from 1.14% in barren unplanted sites to 2.45% in the 10-year-old site. However, it was less than in the few remaining natural mangrove forests (4.02%) in the region. Kairo et al. (2008) compared soil properties of a 12-year-old planted forest with an unplanted site. The planted forest contained more SOM (31%) and silt-clay (38%) than the unplanted sites (22% SOM, 16% silt-clay). Bantilan-Smith et al. (2009) observed higher clay content in natural and restored wetlands than created wetlands, and the natural wetlands had higher silt concentration than restored wetlands.

In some restoration efforts, organic matter amendments are used to help jump-start C and nutrient (N, P) cycling processes. While Stauffer and Brooks (1997) and Bruland and Richardson (2004) found that organic amendments are effective for increasing soil moisture, C, and N and decreasing bulk density in nontidal freshwater wetlands, amendments generally have limited benefits in coastal systems. For example, Gibson et al. (1994) found no increase in SOM when organic amendments (alfalfa, straw) were added to southern California constructed marshes. They attributed this to the permeable, sandy soils which increased decomposition and leaching of N. Likewise, Thompson et al. (1995) reported no benefits of adding organic matter (peat) on nitrogen cycling (nitrification, denitrification) to a restored salt marsh in North Carolina.

Like tidal marshes, for many mangrove restoration projects a major constraint for the development of ecosystem processes is low organic matter content of the planting substrate. McKee and Faulkner (2000) compared soil properties of two sites restored by fill removal and two natural reference mangrove forests in southwestern Florida. The restored sites, 6 and 14 years old, had three to five times less SOM (10%–12%) than the natural forests (38%–56%). Bulk density was three times higher in the restored forest soils. In developing countries, a similar problem exists as planting is done on mudflats that contain little SOM (Al-Kayat and Jones 1999). Osland et al. (2012) compared forest structure and soil properties along a chronosequence of nine mangrove plantings, ranging in age from 1 to 20 years, in Tampa Bay, Florida. There were clear trajectories of increasing soil organic C and N and decreasing bulk density with age for surface (0–10 cm) soils. The 20-year-old site exhibited surface soil properties that were within the range of nine reference mangrove forests.

Development of the benthic infauna community, an important component of tidal wetland food webs, is linked to accumulating SOM (Moy and Levin 1991, Sacco et al. 1994). Craft and Sacco (2003) compared benthic infauna communities along a chronosequence of 1- to 28-year-old constructed tidal marshes. Populations of surface deposit feeders developed quickly whereas subsurface deposit feeders, especially oligochaetes, were slow to develop, requiring up to 20 years to achieve equivalence to infauna communities of natural marshes. Craft and Sacco suggested that, in constructed marshes, 0.5%–2% soil organic C and 500 g m^{-2} of MOM (0–10 cm depth) are needed to produce comparable overall density and diversity of benthic infauna communities.

While many created and restored wetlands may display developmental trajectories that suggest the potential for equivalence with natural sites in the future, there is still a vast deficit in ecosystem services left by the destruction and conversion of wetlands. Thus in order to compensate for the loss of these services successfully on a human time scale, restoration projects must be larger in scale. For instance, Osland et al. (2012) calculated that while a 1:1 replacement area ratio for Florida mangroves wetlands can never make up for previous losses in C burial, a replacement ratio of 2:1 can make up the burial deficit in 19 years. To accomplish ambitious goals for the landscape-level reinstatement of C sequestration, water quality improvement, habitat improvement and the multitude of other ecosystem services provided by tidal wetland soils via restoration will require increasingly large restoration projects. By virtue of their size, these larger projects will require not only the implementation of minimal effort “self-design” concepts but the implementation of these strategies over large geographic scales.

Case Study: Large-Scale Restoration on the Louisiana Gulf Coast

Coastal Louisiana has been shaped over the past 7000 years as the Mississippi River delivered sediments and freshwater to form multiple delta lobes dominated by tidal wetlands, and allowed marsh accretion to keep pace with rising sea levels (Boesch et al. 1994). Increased river-control structures, sea-level rise, and increased oil and gas exploration in the Gulf of Mexico have altered marsh hydrology, leading to marsh subsidence (Boesch et al. 1994, DeLaune et al. 2003, Dahl 2011, Kearney et al. 2011). While there is controversy over the primary cause of wetland loss in Louisiana, most agree that increased sediment delivery would help increase elevation and maintain marsh accretion in the face of rising sea levels.

Freshwater diversions from the Mississippi River have been advocated to maintain or increase marsh surface elevation by increasing accretion and reducing salinity to support a healthy vegetation community (DeLaune et al. 2003, Lane et al. 2006, Day et al. 2009). Crevasse splays, where the levee is deliberately breached, are used to allow river water and sediment to flow into a degraded or subsiding wetland (Boyer et al. 1997, Cahoon et al. 2011). River diversions mimic crevasse splays but rely on engineered structures to control water and sediment inputs (Lane et al. 2006), ideally resulting in large-scale land-building with the addition of freshwater and sediment as have been observed in the Wax Lake Outlet (Figure 18.6).

Lane et al. (2006) followed changes in surface elevation, vertical accretion, and subsidence in marshes along three Mississippi River diversions: Caernarvon, West Point a la Hache, and the Violet river diversions. Of the three sites, the Violet diversion had the lowest discharge and the greatest hydrologic alteration and subsequently lost elevation throughout the study period. The two other diversions had greater accretion and elevation gains which were attributed to greater discharge from the diversions and closer proximity of the sites to the water source. Similarly, DeLaune et al. (2003) tracked changes in marsh accretion through organic matter and mineral deposition in marshes receiving freshwater from the Caernarvon diversion. Sites receiving Mississippi River water had increased sediment deposition, mineral sediment organic matter accumulation, and nutrients. Salinity levels also decreased in response to inputs of Mississippi River water. The authors suggested that river diversions are a sustainable way to restore coastal wetlands and combat land loss through sea-level rise and saltwater intrusion, provided that diversions are appropriately sized.

Some river diversions that have operated for over 20 years in coastal Louisiana to restore coastal marshes have not produced the desired results of increased plant production and soil accretion. In that time, there has been an overall decrease in the amount of vegetated marsh from losses caused by hurricanes like Katrina and Rita, but also from anthropogenic nutrient loading. Kearney et al. (2011) suggest that nutrient-rich river water which was diverted into these systems decreases belowground biomass and increases organic matter decomposition, thereby weakening the stability and resiliency of these wetlands to physical perturbations. Swarzenski et al. (2008) measured porewater and substrate characteristics of organic-rich marshes in coastal Louisiana that have received diverted Mississippi river water for 30 years. They found greater dissolved inorganic nitrogen, phosphorus, and sulfide concentrations, higher pH, and lower dissolved iron in the marshes receiving river water than in reference (no river diversion) marshes. Soils in the restored marshes were more reduced and had a much lower resistance to shear force, making them more susceptible to hurricanes, storms, and tidal action.

Nyman (2014) outlines two ideas regarding river diversions and natural cycles of succession that occur in large river deltas. Sediment diversions mimic the active stage of Delta

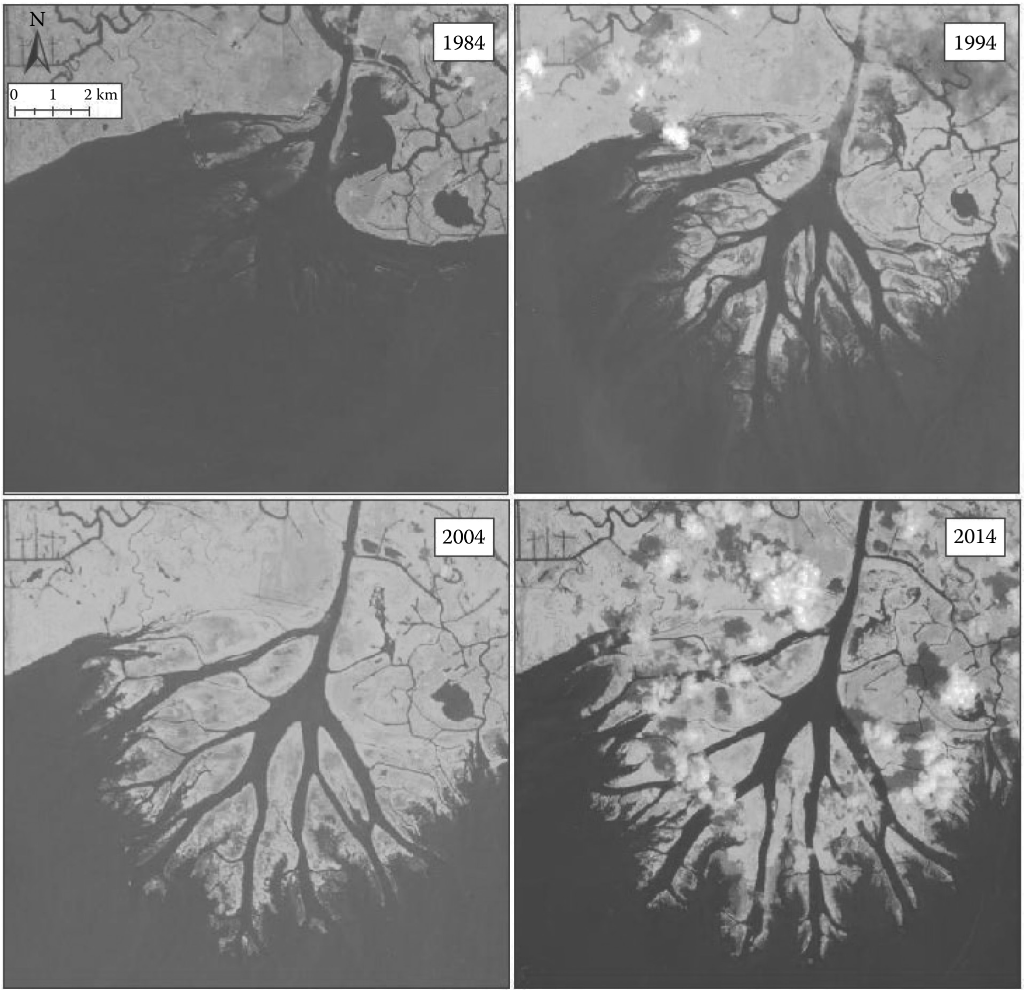


FIGURE 18.6
(**See color insert.**) Development of the Wax Lake Delta, Louisiana, over a 30-year period. The delta formed as the result of an unplanned driver diversion to reduce flooding along the Atchafalaya River in the 1970s. Photos were taken by Landsat 5 (1994–2004) and Landsat 8 (2014) satellites.

Lobe cycle where sediment-rich river water leads to rapid accretion of mineral sediment, creating new emergent wetlands. Freshwater diversions, in contrast, represent the inactive stage of the Delta Lobe Cycle, supplying freshwater to reduce salinity stress over large areas of emergent wetlands, promoting vertical accretion by growth of vegetation and accumulation of plant-derived organic matter. Sediment diversions are designed to build new wetlands whereas freshwater diversions aim to slow the loss of existing wetlands.

The three case studies presented in this text highlight the different challenges, approaches, and expected outcomes of restoration projects (Table 18.3). The benefit of using dredged material is that the planting substrate is already reduced whereas the drawback is that such restorations are relatively small in size. Breaching of dikes and levees relies more on self-design, including providing an adequate sediment supply and creating and restoring tidal wetlands at larger scales. River diversions potentially can restore tidal wetlands

TABLE 18.3

Pros and Cons of Various Methods to Create and Restore Tidal Wetlands

	Pros	Cons
Dredge spoil	Builds new marsh	Relatively small scale
	Readily available supply of sediment	Low nutrient supply (requires fertilization)
	Soil are already reduced	Usually requires planting
Dike breaching	Builds new marsh	Often subsided (may require hydrologic regulation)
	Relative large scale	Requires large external supply of sediment
	Naturally builds tidal channels	
	Relies on natural colonization	
River diversions	Builds new marshes and/or slows loss of existing marsh	Limited to large river deltas
		Nutrient loading may be problematic
	Large scale	
	Relies on natural colonization	

at large spatial scales but may be hindered by water releases that are otherwise used to maintain adequate water levels to support navigation and commerce. Another drawback of river diversions is the potential for eutrophication caused by nutrient-enriched river water that decreases marsh stability and long-term stability (Swarzenski et al. 2008, Kearney et al. 2011).

The Future of Restoration: Challenges

While restoration science has primarily focused on how best to restore tidal wetlands and what limits their restoration, in recent years the long-term viability of coastal wetland restoration projects has come into question (Callaway et al. 2007, Erwin 2008). Rising sea levels, changes in water quality, changes in hydrology and freshwater inputs, and increasing human alteration of the coastal environment are threats to natural, restored, and created wetlands (Erwin 2008). Current rates of sea level rise are 2.2–3.6 mm year^{-1} (Church and White 2011), but are expected to be as much as 15.6 mm year^{-1} by 2100 (Church et al. 2013). As a consequence of global climate change, mean sea level is projected to increase by 0.19–0.83 m by 2100 (Church et al. 2013) with some models projecting increases of >1 m by 2100 (Richardson et al. 2009, Vermeer and Rahmstorf 2009, Rignot et al. 2011). Higher water levels may lead to submergence of tidal wetlands while salt water intrusion will convert tidal fresh marshes and forests to brackish marsh or open water (Craft et al. 2009). Created and restored wetlands often have significant elevation deficits in comparison to natural marshes, either by design (see *Restoring Hydrology* above) or due to subsidence, as in the case of impoundments. At the same time, the presence of levees, dikes, and other structures surrounding restored wetlands, as well as increasing development of the coastal zone, including shoreline armoring and dredging, will limit the ability of wetlands to migrate inland and upstream. The combined effects of sea level rise and urban encroachment will "squeeze" tidal wetlands and severely limit their ability to persist in the future. Today's tidal wetland restoration projects must plan accordingly to ensure that these wetlands have room to migrate inland and upriver.

Hydrologic modifications, particularly the construction of upstream dams and changes in land use, have already been observed to reduce current sediment delivery to the coast (Weston 2014), potentially reducing the ability of created and restored wetlands to accumulate sediments. There is also concern that in sediment limited environments, there is potential for restoration sites to preferentially accumulate sediment over higher elevation natural sites to the determent of natural wetlands (Callaway et al. 2009). Damming and increased freshwater withdrawals will also limit the delivery of freshwater to coastal environments, resulting in elevated salinities in estuaries which can stress coastal marshes and limit plant establishment and productivity and result in marsh dieback, even in salt marshes (Alber et al. 2008, Wieski et al. 2010). Increased evapotranspiration in response to elevated temperatures may further exacerbate salinity stress in wetland environments.

Planning for restoration in the future will need to consider current and future sea level rise, sediment supply, and anthropogenic alterations in sediment delivery. With proper placement, including ability to migrate inland as sea level rises, created and restored tidal wetlands can produce societal benefits for today's and future generations.

References

Alber, M., E. M. Swenson, S. C. Adamowicz, and I. A. Mendelssohn. 2008. Salt marsh dieback: An overview of recent events in the US. *Estuarine, Coastal and Shelf Science* 80: 1–11.

Al-Khayat, J. A. and D. A. Jones. 1999. A comparison of the macrofauna of natural and replanted mangroves in Qatar. *Estuarine, Coastal and Shelf Science* 49 (Supplement A): 55–63.

Alphin, T. D. and M. H. Posey. 2000. Long-term trends in vegetation dominance and infaunal community composition in created marshes. *Wetlands Ecology and Mangement* 8: 317–325.

Anisfeld, S. C. 2012. Biogeochemical responses to tidal restoration. In C. T. Roman, and D. M. Burdick (Eds.) *Tidal Marsh Restoration: A Synthesis of Science and Management.* Island Press, Washington, DC, pp. 39–58.

Baldwin, A. H. and E. F. Derico. 1999. The seed bank of a restored tidal freshwater marsh in Washington, DC. *Urban Ecosystems* 3: 5–20.

Baldwin, A. H. 2004. Restoring complex vegetation in urban settings: The case of tidal freshwater marshes. *Urban Ecosystems* 7: 125–137.

Bantilan-Smith, M., G. Bruland, R. A. MacKenzie, A. R. Henry, and C. R. Ryder. 2009. A comparison of the vegetation and soils of natural, restored, and created coastal lowland wetlands in Hawaii. *Wetlands* 29: 1023–1035.

Boesch, D. F. and R. E. Turner. 1984. Dependence of fishery species on salt marshes: The role of food and refuge. *Estuaries* 7(4A): 460–468.

Boesch, D. F., M. N. Josselyn, A. J. Mehta, J. T. Morris, W. K. Nuttle, C. A. Simenstad, and D. J. P. Swift. 1994. Scientific assessment of coastal wetland loss, restoration and management in Louisiana. *Journal of Coastal Research* SI 20: 1–103.

Boyer, K. E. and J. B. Zedler. 1998. Effects of nitrogen additions on the vertical structure of a constructed cordgrass marsh. *Ecological Applications* 8: 692–705.

Boyer, M. E., J. O. Harris, and R. E. Turner. 1997. Constructed crevasses and land gain in the mississippi river delta. *Restoration Ecology* 7: 85–92.

Brew, D. S. and P. B. Williams. 2010. Predicting the impact of large-scale tidal wetland restoration on morphodynamics and habitat evolution in south San Francisco Bay, California. *Journal of Coastal Research* 265: 912–924.

Broome, S. W., E. D. Seneca, and W. W. Woodhouse, Jr. 1986. Long-term growth and development of transplants of the salt-marsh grass *Spartina alterniflora*. *Estuaries* 9: 63–74.

Broome, S. W., E. D. Seneca, and W. W. Woodhouse, Jr. 1988. Tidal salt marsh restoration. *Aquatic Botany* 32: 1–22.
Bruland, G. L. and C. J. Richardson. 2004. Hydrologic gradients and topsoil additions affect soil properties of Virginia created wetlands. *Soil Science Society of America* 68: 2069–2077.
Cahoon, D. R., D. A. White, and J. C. Lynch. 2011. Sediment infilling and wetland formation dynamics in an active crevasse splay of the Mississippi River Delta. *Geomorphology* 131: 57–68.
Callaway, J. C., G. Sullivan, and J. B. Zedler. 2003. Species-rich plantings increase biomass and nitrogen accumulation in a wetland restoration experiment. *Ecological Applications* 13: 1626–1639.
Callaway, J. C., V. T. Parker, M. C. Vasey, and L .M. Schile. 2007. Emerging issues for the restoration of tidal marsh ecosystems in the context of predicted climate change. *Madrono* 54: 234–248.
Callaway, J. C., V. T. Parker, L. M. Schile, E. R. Herbert, and E. L. Borgnis. 2009. *Dynamics of sediment accumulation in pond A21 at the Island Ponds*. California State Coastal Conservancy, Oakland, CA.
Callaway, J. C., V. T. Parker, M. C. Vasey, L. M. Schile, and E. R. Herbert. 2011. Tidal wetland restoration in San Francisco Bay: History and current issues. *San Francisco Estuary and Watershed Science* 9(3): 1–12.
Cammen, L. M. 1976. Macroinvertebrate colonization of *Spartina* marshes artificially established on dredge spoil. *Estuarine and Coastal Marine Science* 4: 357–372.
Chmura, G. L., S. C. Anisfeld, D. R. Cahoon, and J. C. Lynch. 2003. Global carbon sequestration in tidal, saline wetland soils. *Global Biogeochemical Cycles* 17: 22.1–22.12.
Church, J. A., P. U. Clark, A. Cazenave, J. M. Gregory, S. Jevrejeva, A. Levermann, M. A. Merrifield, G. A. Milne, R. S. Nerem, P. D. Nunn, A. J. Payne, W. T. Pfeffer, D. Stammer, and A.S. Unnikrishnan. 2013. *Sea Level Change*. Cambridge, UK: Cambridge University Press.
Church, J. A. and N. J. White. 2011. Sea-level rise from the late 19th to the early twentry first century. *Surveys in Geophysics* 32: 585–602.
Coleman, J. M., O. K. Huh, and B. DeWitt, Jr. 2008. Wetland loss in world deltas. *Journal of Coastal Research* 24: 1–14.
Craft, C. B. 1997. Dynamics of nitrogen and phosphorus retention during wetland ecosystem succession. *Wetlands Ecology and Management* 4: 177–187.
Craft, C. B. 2000. Co-development of wetland soils and benthic invertebrate communities following salt marsh creation. *Wetlands Ecology and Management* 8: 197–207.
Craft, C. B. 2001. Soil organic carbon, nitrogen and phosphorus as indicators of recovery in restored *Spartina* marshes. *Ecological Restoration* 19: 87–91.
Craft, C. B., S. Broome, and C. Campbell. 2002. Fifteen years of vegetation and soil development after brackish-water marsh creation. *Restoration Ecology* 10: 248–258.
Craft, C. B., S. W. Broome, and E. D. Seneca. 1988. Nitrogen, phosphorus and organic-carbon pools in natural and transplanted marsh soils. *Estuaries* 11: 272–280.
Craft, C. B., E. D. Seneca, and S. W. Broome. 1991. Porewater chemistry of natural and created marsh soils. *Journal of Experimental Marine Biology and Ecology* 152: 187–200.
Craft, C. B., J. Clough, J. Ehma, S. Joye, R. Park, S. Pennings, H. Guo, and M. Machmuller. 2009. Forecasting the effects of accelerated sea-level rise on tidal marsh ecosystem services. *Frontiers in Ecology and the Environment* 7: 73–78.
Craft, C. B., P. Megonigal, S. Broome, J. Stevenson, R. Freese, J. Cornell, L. Zheng, and L. Sacco. 2003. The pace of ecosystem development of constructed *Spartina alterniflora* marshes. *Ecological Applications* 13: 1417–1432.
Craft, C. B., J. Reader, J. N. Sacco, and S. W. Broome. 1999. Twenty-five years of ecosystem development of constructed *Spartina alterniflora* (Loisel) marshes. *Ecological Applications* 9: 1405–1419.
Craft, C. B. and J. N. Sacco. 2003. Long-term succession of benthic infauna communities on constructed *Spartina alterniflora* marshes. *Marine Ecology Progress Series* 257: 45–58.
Crooks, S., D. Herr, J. Tamelander, D. Laffoley, and J. Vandever. 2011. Mitigating climate change through restoration and management of coastal wetlands and near-shore marine ecosystems: Challenges and opportunities. *Environment Department Paper 121*, World Bank, Washington, DC.
Cui, B., Q. Yang, Z. Yang, and K. Zhang. 2008 Evaluating the ecological performance of wetland restoration in the Yellow River Delta, China. *Ecological Engineering* 35: 1090–1103.

Dahl, T. E. 2011. *Status and Trends of Wetlands in The Conterminous United States 2004 to 2009*. U.S. Department of Interior, Fish and Wildlife Service, Washington, DC.

Day, J. W. Jr., L. D. Britsch, S. R. Hawes, G. P. Shaffer, D. J. Reed, and D. Cahoon. 2000. Pattern and process of land loss in the Mississippi Delta: A spatial and temporal analysis of wetland habitat change. *Estuaries* 23: 425–438.

Day, J. W., J. E. Cable, J. W. Cowan, Jr., R. DeLaune, K. de Mutsert, B. Fry, H. Mashriqui, D. Justic, P. Kemp, R.R. Lane, J. Rick, S. Rick, L. P. Rozas, G. Snedden, E. Swenson, R.R. Twilley, and B. Wissel. 2009. The impacts of pulsed reintroduction of river water on a Mississippi Delta coastal basin. *Journal of Coastal Research* 54: 225–243.

DeLaune, R. D., A. Jugsujinda, G. W. Peterson, and W. H. Patrick, Jr. 2003. Impact of Mississippi River freshwater reintroduction on enhancing marsh accretionary processes in a Louisiana estuary. *Estuarine, Coastal and Shelf Science* 58: 653–662.

Edwards, K. R. and C.E. Proffitt. 2003. Comparison of wetland structural characteristics between created and natural salt marshes in southwest Louisiana, USA. *Wetlands* 23: 344–356.

Eertman, R. H. M., B. W. Kornman, E. Stikvoort, and H. Verbeek. 2002. Restoration of the Sieperda Tidal Marsh in the Scheldt Estuary, The Netherlands. *Restoration Ecology* 10: 438–449.

Erwin, K. L. 2008. Wetlands and global climate change: The role of wetland restoration in a changing world. *Wetlands Ecology and Management* 17: 71–84.

Duke, N. C. 1996. Mangrove reforestation in Panama: An evaluation of planting in areas deforested by large oil spill. In *Restoration of mangrove ecosystems*, ed. C.D. Field, 209–232. Okinawa, Japan: International Society for Mangrove Ecosystems.

Duke, N. C., J. O. Meynecke, S. Dittman, A. M. Ellison, K. Anger, U. Berger, S. Cannicci, K. Diele, K.C. Ewel, C.D. Field, N. Koedam, S.Y. Lee, C. Marchland, I. Norhaus, and F. Dahdouh-Guebas. 2007. A world without mangroves? *Science* 317: 41–42.

Edwards, K. R. and C. E. Proffitt. 2003. Comparison of wetland structural characteristics between created and natural salt marshes in southwest Louisiana, USA. *Wetlands* 23: 344–356.

Ellison, A. M. 2000. Mangrove restoration: Do we know enough? *Restoration Ecology* 8: 219–229.

Engle, V. D. 2011. Estimating the provision of ecosystem services by Gulf of Mexico coastal wetlands. *Wetlands* 31: 179–193.

Esteves, L. S. 2014. What is managed realignment? In *Managed realignment: A viable long-term coastal management strategy?*, ed. L.S. Esteves, 19–31. Springer Briefs in Environmental Science. The Netherlands: Springer.

Fearnley S. 2008. The soil physical and chemical properties of restored and natural back-barrier salt marshes on Isles Dernieres, Louisiana. *Journal of Coastal Research* 24: 84–94.

Field, C. D. 1996. *Restoration of mangrove ecosystems*. Okinawa, Japan: International Society for Mangrove Ecosystems.

Food and Agricultural Organization. 2007. *The world's mangroves: 1980-2005*. Rome, Italy: Food and Agricultural Organization of the United Nations.

Ford, M. A., D. R. Cahoon, and J. C. Lynch. 1999. Restored salt marsh elevation in a rapidly subsiding salt marsh by thin-layer deposition of dredged material. *Ecological Engineering* 12: 189–205.

Gibson, K. D. J. B. Zedler, R. Langis. 1994. Limited response of cordgrass (Spartina foliosa) to soil amendments in a constructed marsh. *Ecological Applications* 4: 757–767.

Harvey, H. T., P. B. Williams, and J. Haltiner. 1982. Guidelines for enhancement and restoration of diked historic baylands. Technical report, San Francisco: Madrone Associates and Bay Conservation and Development Commission.

Havens, K. J., L. M. Varnell, and B. D. Watts. 2002. Maturation of a constructed tidal marsh relative to two natural reference marshes over 12 years. *Ecological Engineering* 18: 305–315.

Imbert, D., A. Rousteau, and P. Scherrer. 2000. Ecology of mangrove growth and recovery in the Lesser Antilles: State of knowledge and basis for restoration projects. *Restoration Ecology* 8: 230–236.

Kairo, J. G., J. K. S. Lang'at, F. Dahdouh-Guebas, J. Bosire, and M. Karachi. 2008. Structural development and productivity of replanted mangrove plantations in Kenya. *Forest Ecology and Management* 255: 2670–2677.

Kaly, U. L. and G. P. Jones. 1998. Mangrove restoration: A potential tool for coastal management in tropical developing countries. *Ambio* 27: 656–661.
Kearney, M. S., J. C. A. Riter, and R. E. Turner. 2011. Freshwater river diversions for marsh restoration in Louisiana: Twenty-six years of changing vegetative cover and marsh area. *Geophysical Research Letters* 38: LI6405.
Kelly, N. M. 2001. Changes to the landscape pattern of coastal North Carolina wetlands under the Clean Water Act, 1984–1992. *Landscape Ecology* 16: 3–16.
Ket, W. A., J. P. Schubauer, and C. B. Craft· 2011. Effects of five years of nitrogen and phosphorus additions on a *Zizaniopsis miliacea* tidal freshwater marsh. *Aquatic Botany* 95: 17–23
Kusler, J. A. and M. E. Kentula. 1990. *Wetland creation and restoration: The status of the science*. Washington, DC: Island Press.
Lane, R. R., J. W. Day, Jr., and J. N. Day. 2006. Wetland surface elevation, vertical accretion, and subsidence at three Louisiana estuaries receiving diverted Mississippi river water. *Wetlands* 26: 1130–1142.
Langis, R., M. Zalejko, and J. B. Zedler. 1991. Nitrogen assessments in a constructed and a natural salt marsh of San Diego Bay. *Ecological Applications* 1: 40–51.
Leck, M. A. 2003. Seed-bank and vegetation development in a created tidal freshwater wetland on the Delware River, Trenton, New Jersey, USA. *Wetlands* 23: 310–343.
Levin, L. A., D. Talley, and G. Thayer. 1996. Succession of macrobenthos in a created salt marsh. *Marine Ecology Progress Series* 141: 67–82.
Lewis, R. R. III. 2000. Ecologically based goal setting in mangrove forest and tidal marsh restoration. *Ecological Engineering* 15: 191–198.
Lindau, C. W. and L. R. Hossner. 1981. Substrate characteristics of an experimental marsh and three natural marshes. *Soil Science Society of Journal* 45: 1171–1176.
Marcus, L. 2000. Restoring tidal wetlands at Sonoma Baylands, San Francisco Bay, California. *Ecological Engineering* 15: 373–383.
McKee, K. L. and P. L. Faulkner. 2000. Restoration of biogeochemical function in mangrove forests. *Restoration Ecology* 8: 247–259.
Mendelssohn, I. A. and N. L. Kuhn. 2003. Sediment subsidy: Effects on soil-plant responses in a rapidly submerging coastal salt marsh. *Ecological Engineering* 21: 115–128.
Miller, R. L., M. Fram, R. Fujii, and G. Wheeler. 2008. Subsidence reversal in re-established wetlands in the Sacramento-San Joaquin Delta, California, USA. *San Francisco Estuary and Watershed Science* 6: 1–20.
Morgan, P. A. and F. C. Short. 2002. Using functional trajectories to track constructed salt marsh development in the Great Bay Estuary, Maine/New Hampshire, U.S.A. *Restoration Ecology* 10: 461–473.
Morris, J. T., P. V. Sundareshwar, C. T. Nietch, B. Kjerfve, and D. R. Cahoon. 2002. Response of coastal wetlands to rising sea level. *Ecological Applications* 83: 2869–2877.
Moy, L. D. and L. A. Levin. 1991. Are *Spartina* marshes a replaceable resource? A functional approach to evaluation of marsh creation efforts. *Estuaries* 14: 1–16.
Nyman, J. A. 2014. Integrating successional ecology and the Delta Lobe Cyles in wetland research and restoration. *Estuaries* 37: 1490–1505.
Orr, M., S. Crooks, and P. B. Williams. 2003. Will restored tidal marshes be sustainable? In *Issues in San Francisco Estuary tidal wetlands restoration*, ed. L.R. Brown. San Francisco: San Francisco Estuary and Watershed Science.
Osgood, D. and J. Zieman. 1993. Spatial and temporal patterns of substrate physicochemical parameters in different-aged barrier island marshes. *Estuarine, Coastal and Shelf Science* 37: 421–436.
Osland, M. J. et al. 2012. Ecosystem development after mangrove wetland creation: Plant-soil change across a 20-year chronosequence. *Ecosystems* 15: 848–868.
Peterson, G. W. and R. E. Turner. 1994. The value of salt marsh edge vs interior as habitat for fish and decapod crustaceans in a Louisiana tidal marsh. *Estuaries* 17: 235–262.
Pethick, J. 2002. Estuarine and tidal wetland restoration in the United Kingdom: Policy versus Practice. *Restoration Ecology* 10: 431–437.

Portnoy, J. W. 1999. Salt marsh diking and restoration: biogeochemical implications of altered wetland hydrology. *Environmental Management* 24: 111–120.

Posey, M. H., T. D. Alphin, and C. M. Powell. 1997. Plant and infaunal communities associated with a created marsh. *Estuaries* 20: 42–47.

Redfield, A. C. 1972. Development of a New England salt marsh. *Ecological Monographs* 42: 201–237.

Richardson, K. et al. 2009. Synthesis Report, Climate Change—Global risks, challenges and decisions. Copenhagen, Denmark.

Rignot, E., I. Velicogna, M. Van den Broeke, A. Monaghan, and J. Lenaerts. 2011. Acceleration of the contribution of the Greenland and Antarctic ice sheets to sea level rise. *Geophysical Research Letters* 38: L05503.

Reed, D. J. 2002. Understanding tidal marsh sedimentation in the Sacramento-San Joaquin delta, California. *Journal of Coastal Research* 36: 605–611.

Ren, H., S. Jian, H. Lu, Q. Zhang, W. Shen, W. Han, Z. Yin, and Q. Guo. 2008. Restoration of mangrove plantations and colonization by native species in Leizhou Bay, South China. *Ecological Restoration* 23: 401–407.

Sacco, J. N., E. D. Seneca, and S. W. Broome. 1994. Infauna community development of artificially established salt marshes in North Carolina. *Estuaries* 17: 489–500.

Saenger, P. 2002. *Mangrove ecology, silviculture and conservation*. Dordrecht, The Netherlands: Kluwer Academic Publishers.

Schrift, A. M., I. A. Mendelssohn, and M. D. Materne. 2008. Salt marsh restoration with sediment-slurry amendments following a drought-induced large-scale disturbance. *Wetlands* 28: 1071–1085.

Simenstad, C. A. and R. M. Thom. 1996. Functional equivlancy of the Gog-Le-Hi-Te estuarine wetland. *Ecological Applications* 6: 38–56.

Sklar, F. H. et al. 2005. The ecological-societal underpinnings of Everglades restoration. *Frontiers in Ecology and the Environment* 3: 161–169.

Stauffer, A. L. and R. P. Brooks. 1997. Plant and soil responses to salvaged marsh surface and organic matter amendments at a created wetland in central Pennsylvania. *Wetlands* 17: 90–105.

Streever, W. J. 2000. *Spartina alterniflora* marshes on dredged material: A critical review of the ongoing debate over success. *Wetlands Ecology and Management* 8: 295–316.

Swarzenski, C. M., T. W. Doyle, B. Fry, and T. G. Hargis. 2008. Biogeochemical response of organic-rich freshwater marshes in the Louisiana deltaic plain to chronic river water influx. *Biogeochemistry* 90: 49–63.

Thompson, S. P., H. W. Paerl, and M. C. Go. 1995. Seasonal patterns of nitrification and denitrification in a natural and a restored salt marsh. *Estuaries* 18: 339–408.

Turner, R. E., E. M. Swenson, and J. M. Lee. 1994. A rationale for coastal wetland restoration through spoil bank management in Louisiana, USA. *Environmental Management* 18(2): 271–282.

Turner, R.E. 1997. Wetland loss in the northern Gulf of Mexico: multiple working hypotheses. *Estuaries* 20(1): 1–13.

Valiela, I., J. L. Bowen, and J. K. York. 2001. Mangrove forests: One of the world's threatened ecosystems. *BioScience* 51: 807–815.

Valiela, I. and J. M. Teal. 1974. Nutrient limitation in salt marsh vegetation. In *Ecology of Halophytes*, eds. R. J. Reidmold and W. H. Queen. Academic Press. pp. 547–563.

Vandenbruwaene, W., T. Maris, T. J. S. Cox, D. R. Cahoon, P. Meire, and S. Temmerman. 2011. Sedimentation and response to sea-level rise of a restored marsh with reduced tidal exchange: Comparison with a natural tidal marsh. *Geomorphology* 130: 115–126.

Van Staveren, M. F., J. F. Warner, J. P. M. van Tatenhove, and P. Wester. 2014. Let's bring in the floods: de-poldering in the Netherlands as a strategy for long-term delta-survival? *Water International* 39(5): 686–700.

Vermeer, M. and S. Rahmstorf. 2009. Global sea level linked to global temperature. *Proceedings of the National Academy of Sciences of the United States of America* 106: 21527–21532.

Warren, R. S., P. E. Fell, R. Rozsa, A. H. Brawley, A. C. Orsted, E. T. Olson, V. Swamy, and W. A. Niering. 2002. Salt marsh restoration in Connecticut: 20 years of science and management. *Restoration Ecology* 10(3): 497–513.

Weston, N. B. 2014. Declining sediments and rising seas: an unfortunate convergence for tidal wetlands. *Estuaries and Coasts* 37: 1–23.
Wieski, K., H. Guo, C. B. Craft, and S. C. Pennings. 2010. Ecosystem functions of tidal fresh, brackish, and salt marshes on the Georgia coast. *Estuaries and Coasts* 33: 161–169.
Williams, P. B. and M. K. Orr. 2002. Physical evolution of restored breached levee salt marshes in the San Francisco Bay estuary. *Restoration Ecology* 3: 527–542.
Woodhouse, W. W. Jr. and P. L. Knutson. 1982. Atlantic coastal marshes. In *Creation and restoration of coastal plant communities*, ed. R.R. Lewis, III, 45–70. Boca Raton, FL: CRC Press.
Zedler, J. B. and J. C. Callaway. 1999. Tracking wetland restoration: Do mitigation sites follow desired trajectories? *Restoration Ecology* 7(1): 69–73.
Zedler, J. B. and J. C. Callaway. 2000. Evaluating the progress of engineered tidal wetlands. *Ecological Engineering* 15: 211–225.
Zedler, J. B., J. C. Callaway, and G. Sullivan. 2001. Declining biodiversity: Why species matter and how their functions might be restored in Californian tidal marshes. *BioScience* 51: 1005–1017.

19

Soil Restoration: The Foundation of Successful Wetland Reestablishment

Curtis J. Richardson, Gregory L. Bruland, Matthew F. Hanchey, and Ariana E. Sutton-Grier

CONTENTS

Introduction

Scientific advances in our understanding of the functional importance of wetland ecosystems in providing key ecosystem services on the landscape has led to changes in federal policies aimed at protecting wetlands over the past three decades. These policies include Section 404 of the Clean Water Act (1972), the "no net loss" policy (1992), and the U.S. Army Corps of Engineers and Environmental Protection Agency compensatory mitigation rule (2002, USACE and USEPA 2008) and requirements for wetlands that are destroyed by development activities. The "no net loss" policy sought to replace lost wetland habitat with new habitat by restoring and/or creating wetlands, and is now the cornerstone of wetland conservation in the United States (Mitsch and Gosselink 2007). As a result, numerous federal and state agencies, non-governmental organizations, and private landowners are engaged in wetland restoration and creation across the United States with a keen focus on getting the proper hydrologic conditions needed to support each wetland type's native vegetation but often with little guidance or thought about the importance of creating the soil characteristics needed to sustain the restoration. Soils are the foundation of all terrestrial ecosystems, including wetlands. Soil conditions influence many ecosystem processes and properties, including plant productivity and diversity, microbial community activity and diversity, biogeochemical cycles, and hydrology (Figure 19.1). Here we review these relationships in detail. The goal of this chapter is to quantify how restoring or maintaining physical and chemical soil characteristics of restored wetlands, including

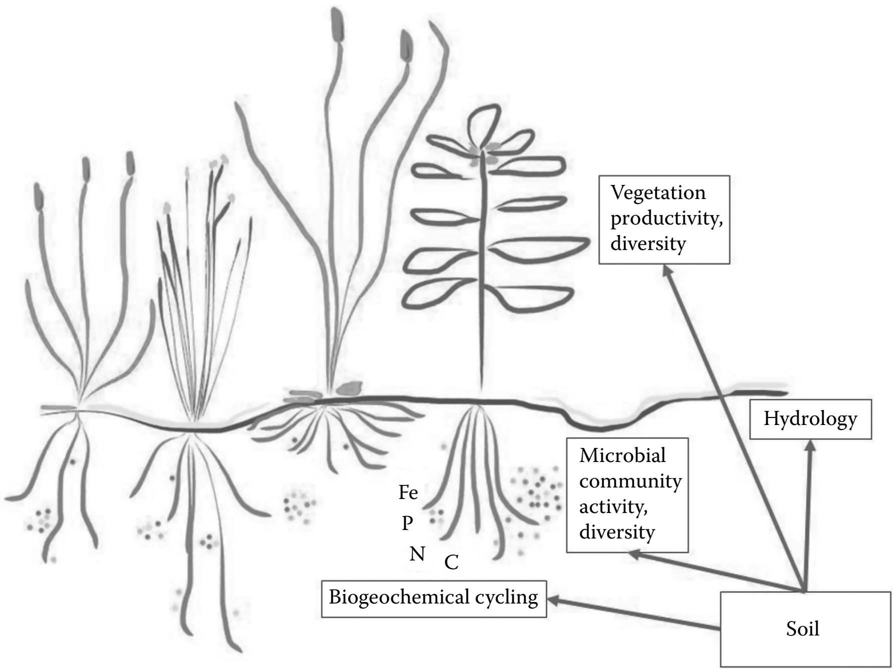


FIGURE 19.1
Conceptual diagram of soils as the foundation of a restored wetland. Soil properties and processes can influence plant diversity and productivity, microbial diversity and activity, biogeochemical cycles, and hydrology.

spatial and microtopographic features and organic matter, results in a more successful and sustainable wetland restoration.

New strategies for accomplishing restoration include manipulating the field site to ensure appropriate wetland hydroperiod and hydropattern to promote the development of hydric soil properties, planting of representative wetland vegetation to hasten the return of a target wetland plant community, and adding soil or organic amendments to the current soil to provide nutrients and enhance successional patterns. The outcome of the restoration process will depend on the interactions between hydrologic restoration strategies and the physical and chemical properties of the site prior to restoration, especially the soils (Shaffer and Ernst 1999; Tweedy and Evans 2001; Unghire et al. 2011). Thus, a strategy that is successful at one site may not have the same results in a site with different hydrologic characteristics, drainage patterns, or soil characteristics. Likewise, a single restoration plan may not produce the same results at all points across a site if that site contains significant internal soil heterogeneity.

Soils are characterized by a high degree of spatial variability due to the combined action of physical, chemical, or biological processes that operate with different intensities and at different scales (Goovaerts 1998; Fennessy and Mitsch 2001; Bruland et al. 2006, 2009). In wetlands, these processes include surface runoff, erosion, overbank flooding, sediment deposition, groundwater inputs, fire, animal burrowing, litter production, and root activity. Consequently, studies of wetland soils should attempt to quantify spatial variability (Stolt et al. 2000; Johnston et al. 2001). Sampling designs that quantify such variability, however, are seldom done for two main reasons: (1) the difficulty of establishing spatially-explicit sampling designs at remote, wet, and often densely vegetated sites; and (2) the need to collect a large number of samples for adequate spatial coverage (Bridgham et al.

2001). The few spatially explicit studies that have been conducted in natural wetlands have found that soil properties exhibited high spatial variability (Lyons et al. 1998; Johnston et al. 2001; Hanchey 2002; Bruland and Richardson 2004, 2005a).

One of the clearest manifestations of internal heterogeneity in natural wetlands is in patterns of vegetation. Vegetative pattern results from such factors as dispersal, germination, and competition; but importantly, community structure also directly reflects soil heterogeneity. In wetlands, the expression of these factors is also strongly affected by patterns of flooding. Fluctuating water levels can play a critical role in determining community composition. Certain wetland species may germinate under flooded conditions, while others can only establish during drawdowns (Casanova and Brock 2000). Individuals that become established during dry periods may have variable responses to renewed flooding, and some species that can establish only during dry periods grow well under flooded soil conditions (Casanova and Brock 2000; Osland et al. 2011), while in others, the onset of flooding can alter productivity and biomass allocation (Lenssen et al. 1999). For example, flooding, and the concomitant low soil redox potentials, can inhibit uptake of nutrients of flood intolerant species, while those species adapted to flooded conditions remain unaffected (Pezeshki et al. 1999).

When patterns of flooding are not equal across a wetland, spatial heterogeneity can arise (Seabloom et al. 2001). This may be along an elevation gradient such as adjacent to a river where elevated soil berms often form. In bottomland hardwood forests, small changes in soil elevation, and by extension flooding parameters, can lead to changes in stand composition and structure (Bledsoe and Shear 2000). Alternatively, in areas where soil microtopography is such that points of unequal elevation are intermixed, vegetation may be patchy (Grace et al. 2000); this heterogeneity serves to promote diversity within a wetland by allowing species with different environmental requirements to coexist in close association (Grace et al. 2000; Hanchey 2002).

Spatial heterogeneity in soil properties has been observed in many natural systems as noted earlier, though it has been the subject of relatively few restoration studies. As is the case for vegetation, many soil properties are correlated with elevation (Reese and Moorhead 1996; Stolt et al. 2001). For example, soil texture, which can have an effect on many other soil properties, is, in some cases, finer (i.e., a greater proportion of clay-sized particles to sand-sized particles) at lower elevations (Pachepsky et al. 2001). Two processes may explain this pattern, depending on the geomorphology of the area in question: (1) in alluvial wetlands, the deposition of coarse particles close to the river occurs during flood events (Stolt et al. 2001); and, (2) in systems fed by rainfall, the downward movement of fine particles occurs during rain events (Pachepsky et al. 2001). Organic carbon (C) and nitrogen (N) may increase with decreasing elevation due to more frequent flooding and decreased decomposition rates at low spots on the landscape (Stolt et al. 2001). Studying geomorphic position rather than elevation in alluvial wetlands may be more important as Johnston et al. (2001) and Sutton-Grier et al. (2009) found that patterns of soil nutrients vary with geomorphic position: higher elevation zones contained higher concentrations of nitrate, probably due to high rates of nitrification in more oxidized soils.

Therefore, if spatial patterns of plant and soil properties are not explicitly acknowledged in a restoration plan, numerous problems may arise, depending on site conditions. First, land use prior to restoration may have contributed to the homogenization of the site, as happens in cultivated fields (Paz-Gonzalez et al. 2000). In this case, the restored wetland may be less diverse in its species composition (Grace et al. 2000; Bruland et al. 2003), and it may lack the functional benefits that arise from varying soil properties and plant species. Second, if the restored wetland remains heterogeneous, but this characteristic has not been anticipated during sampling, assessment metrics for the site may be misleading. For example, when taking multiple samples across the site and calculating the average soil C accumulation, the average will be only a partial descriptor of C dynamics across the site. Third, without the use of spatial soils data from reference wetlands (i.e., least disturbed wetlands in the region or watershed on similar soil types) there is little information upon which to design the amount of microtopography needed to maintain vegetation patterns, or the necessary nutrient or physical soil conditions needed to sustain species growth and plant distributions.

Today many researchers use natural wetlands (NWs) as reference sites to assess success of created (CWs) and restored wetlands (RWs) (Bishel-Machung et al. 1996; Balcombe et al. 2005; Richardson et al. 2011). The use of reference wetlands is based on the underlying assumptions that intact NWs exhibit high ecological function, and wetlands sharing similar hydrologic characteristics, vegetation communities, and soil properties (spatial heterogeneity, soil C content, and microtopography, etc.) will function similarly (Brinson and Rheinhardt 1996; Stolt et al. 2000; Zampella and Laidig 2003). Kentula (2000) recommended comparing identical ecological parameters among populations of CWs, RWs, and NWs within a region to extrapolate results beyond site-specific studies. Importantly, two of these parameters, vegetation and soil characteristics are often directly correlated, relatively easy to sample and are involved in complex interactions that contribute to wetland function (Craft et al. 1988; Shaffer and Ernst 1999; Bruland et al. 2006; Richardson 2008).

Soil properties of CWs and RWs have likewise been shown to differ from those of NWs, especially during the first few years of restoration (Bishel-Machung et al. 1996; Bruland et al. 2003; Bruland and Richardson 2005a, b, 2006; Unghire et al. 2011). This is problematic as soils are the physical foundation for every wetland ecosystem and both plants and animals depend on wetland soils for growth and survival (Stolt et al. 2000; Bruland and Richardson 2004). These edaphic differences are the result of a variety of factors. First, the removal of topsoil during creation or restoration of a wetland results in disturbance of soils (Shaffer and Ernst 1999) and can lower concentrations of soil organic matter (SOM). Differences in SOM can significantly affect many other soil properties, such as total-percent N, bulk density (Db), and pH (Bishel-Machung et al. 1996; Sutton-Grier et al. 2009; Unghire et al. 2011). Second, the use of heavy machinery results in soil compaction, increasing Db in CWs and RWs. Also, differences in hydrology between CWs, RWs, and NWs can also affect soil properties (Craft et al. 2002). For example, when the hydroperiod is lengthened, anaerobic conditions slow decomposition rates and allow for organic matter to accumulate in the soil, decreasing Db and pH (Craft et al. 2002).

Collectively, the aforementioned studies suggest that several key aspects of soil characteristics must be understood and incorporated into modern wetland restoration in order to maintain the optimal successional trajectory that will ensure continued wetland ecosystem functions and sustainability. Restoration characteristics that should be included when considering reestablishment of wetland soil properties are as follows: (1) spatial variability in soil characteristics as it relates to restoration of ecosystem processes and vegetation patterns; (2) microtopography and its role in ecosystem function and plant survival; (3) the role of soil organic matter and organic amendments to plant survival and ecosystem functions; (4) the effect of soil texture, bulk density, and compaction on vegetation responses; and (5) soil nutrient concentrations and their availability in soils. To help assess and quantify the role of these key factors, we draw upon several restoration case studies focused on the function of these essential soil characteristics in determining the successful return of ecosystem structure and biogeochemical processes. A detailed description of soil potential soil biogeochemical indicators for assessing wetland impacts and restoration success has been developed by Reddy and DeLaune (2008) and is characterized by Fennessy and Wardrop in Chapter 17 of this volume and is not covered here. Soil indicators range from basic measurements like soil Db or pH to more in-depth indicators requiring more intense field and laboratory methods like extractable ions or microbial biomass to intensive biological and physicochemical measures like soil sorption coefficients or denitrification rates.

Soil Characteristics

Spatial Variability of Soils in Natural versus Restored Wetlands

One of the first studies to thoroughly look at the influence of spatial soil differences in a restored wetland as it relates to vegetation was done by Hanchey (2002) and Bruland et al. (2003) in a Carolina bay (Figure 19.2). Bays are shallow, non-alluvial, depressional wetlands of unknown origin, which vary in size from a few hectares to over 1000 ha, and over 5000 of these bays are located along the southeastern coastal plain of the United States (Sharitz and Gibbons 1982; Richardson and Gibbons 1993; Richardson 2012). Carolina bays have several different characteristic community types, including tree-dominated, shrub-dominated, and herbaceous-dominated wetlands (Poiani and Dixon 1995; Kirkman et al. 2000). Coarse variability in soil properties of Carolina bays is a well-recognized characteristic, due to the common pattern of a sandy soil around the rim, with either a clay or peat

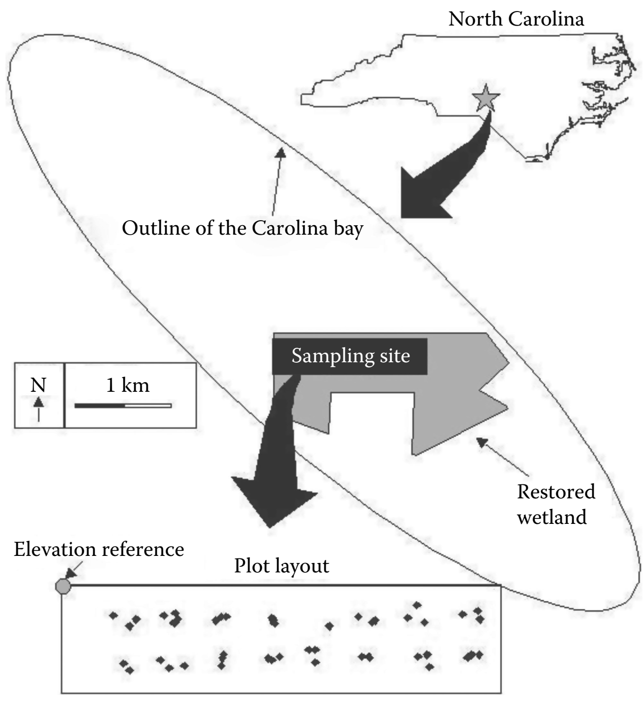


FIGURE 19.2
Site map showing the location of the Carolina bay wetland, the position of the restored wetland within the bay, and the arrangement of sample plots within the restored wetland.

substrate in the interior of the bays (Sharitz and Gibbons 1982). Fine-scale variation has also been observed: Reese and Moorhead (1996) sampled a bay at 10 cm elevation intervals and determined that as elevation decreased across the wetland organic C, clay content, cation exchange capacity, and base saturation increased. Thus a spatial analysis of the soils is critical to aid in the selection of appropriate vegetation across a bay restoration.

The restoration study was done at the Barra Farms Cape Fear Regional Mitigation Bank, a 250 ha restored wetland in the coastal plain of North Carolina, to determine the influence of elevation, soils, and hydrology on plant vegetation (Figure 19.2). The restored wetland occupied a portion of an 800 ha Carolina bay wetland (Hanchey 2002). The majority of the bay was cleared and drained for farming in the mid-1960s, and portions of the bay surrounding the restored wetland remain in agriculture. Restoration consisted of filling 3300 m of ditches and planting of 192,000 bare-root tree seedlings. Seedlings were planted at a density of approximately 1090 stems per ha. Heterogeneity was introduced into the restoration process in two ways. First, microtopography was introduced by scarring the soil in random patterns with a tractor. Second, two zones were established within the restored wetland: a nonalluvial swamp forest zone and a wet hardwood forest zone. Within these zones the mixture of species planted was slightly different, primarily in that species such as *Quercus* spp. were planted more frequently in the wet hardwood forest zone, while *Taxodium* spp. and *Nyssa* spp. were planted at a higher frequency in the swamp forest zone. A full analysis including detailed methods of the spatial sampling design used to determine soil and hydrologic characteristics and vegetation patterns can be found in Hanchey (2002) and Bruland et al. (2003). A brief summary of their key findings is given below to provide insights into the importance of understanding spatial soil patterns when considering vegetation plantings and ecosystem functions.

Elevation, Soils, and Vegetation Data

The Barra Farms site contained a change in elevation of approximately 6 m. Referenced to the northwest corner of the site, elevation generally decreased to the south and east (Figure 19.3a). Slope across the site was approximately 0.5%. Soil sand content ranged from 96.1% at the western end of the site to 55.7% at the eastern end of the site, with a fairly smooth decline from west to east (Figure 19.3b). Total C ranged from 2.0% at the western end of the site to 33.9% at the eastern end; however, distribution was generally patchy across the site (Figure 19.3c). Total TN (0.08%–1.15%) followed a pattern similar to that of carbon (Data not shown). Like C, extractable nitrate (NO_3-N) (1.09–32.94 μg/g) generally increased from east to west, while extractable ammonium (NH_4-N) (2.28–10.95 μg/g) decreased over that distance. Extractable phosphorus (PO_4-P) (7.5–37.97 μg/g) was patchy across the site and did not show a distinct trend (Figure 19.3d). Cations, which were also patchily distributed, ranged as follows: Na- 0.01–0.02, Mg- 0.09–1.05, Ca- 0.56–5.98, and K- 0.01–0.17 mg/g.

Three contrasting patterns of abundance of the most common species encountered at the study site follow the soil variability to some degree (Figure 19.4a–d). Of the volunteer species, the species most frequently encountered (i.e., present in the largest number of plots) was *Scirpus cyperinus* (OBL), which occurred in 83% of the 47 plots taken. *Scirpus* was most common on the eastern side of the study site with the lowest sand and highest C content (Figure 19.4a); however, it was present to some extent throughout the entire site. In some plots in the eastern portion of the site, *Scirpus* occurred in dense stands, occupying 85%–100% of the plot. The second most common species encountered was *Solidago altissima* (FACU, a non-wetland plant species), which occurred in 53% of the plots; *Solidago* was commonly encountered in the sandy western portion of the site but was entirely absent

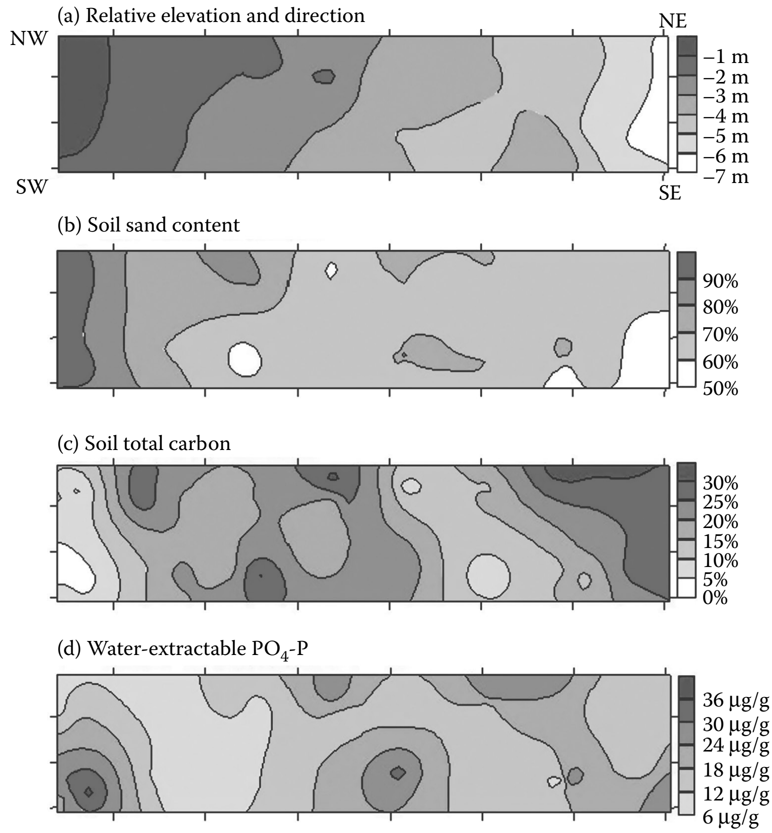


FIGURE 19.3
(See color insert.) (a–d) Distribution of selected soil parameters across the restored Carolina bay wetland in Cumberland Co., NC. Elevation measurements are referenced to the northwest (upper left) corner.

from the wetter eastern portion (Figure 19.4b). The third most common species, *Juncus dichotomus* (FACW), was found in 21 plots, and reached its highest cover values in the southeastern portion of the site; a map is not presented due to the similarity of the distribution of *Juncus* with that of *Solidago*. The fourth most abundant species, *Polygonum punctatum* (FACW), found in 19 plots, generally reached its highest cover values in the central portion of the site, which was the wettest area with high soil carbon (Figure 19.4c). *Taxodium distichum* was the most frequently observed planted species, found in 15 of the 47 plots, mostly in the wetter and lower areas (Figure 19.4d).

Patterns of soil and vegetation at this site have several important implications for the way in which success of the restoration can be evaluated. First, soil heterogeneity suggests that, when compared to nearby reference sites, not all parts of this site would be considered equally successful. The western portion of the site is drier, sandier, and contains more upland species than the eastern portion of the site; the most successful examples of planted species are found in the eastern portion as well. Second, it is likely that the ecosystem functions of interest in this wetland are not present at equivalent levels across the site. For example, coupled nitrification/denitrification is integral in determining a wetland's capacity to improve water quality. Patterns of nitrate and ammonium, which differed from west to east across the site, as noted earlier, suggest that nitrification and

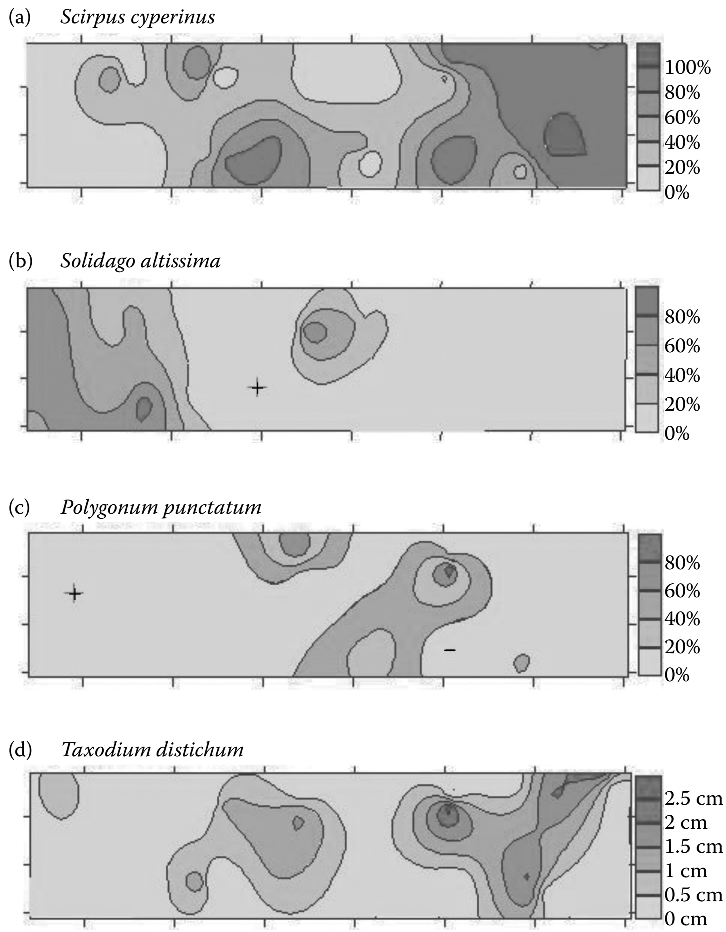


FIGURE 19.4
(See color insert.) (a–c) Distribution of the three most commonly observed species at the restored Carolina bay wetland in Cumberland Co., NC. (d) Distribution of planted *Taxodium distichum* saplings.

denitrification are acting in very different ways at the sandier and drier western areas (Figure 19.3a and b) than they are on the eastern lowland areas, suggesting that the effect of this wetland on water quality would be variable. Third, current properties of the bay can affect its long-term development. For example, soil texture contributes to organic matter accumulation because clays protect organic matter from decomposition (Hassink 1995). The predominance of sandy soil at higher elevations at this site suggests they will accumulate organic matter at rates slower than the lower elevations (Figure 19.3a and c). Correlations between sand and C and N also suggest this effect is already in place, and it may become more exaggerated with time (Hanchey 2002). Also, Carolina bays have several potential stable vegetation states, influenced in part by hydrology and soil characteristics (Kirkman et al. 2000); heterogeneity across the bay has led to trends in early successional vegetation, which has resulted in much different long-term successional patterns across the site. For example, wetter sites have become dominated by *Taxodium* and *Nyssa*, as planned in the restoration; in contrast, the higher and drier portion of the site (Figure 19.3a) have become dominated by volunteer tree species that perform well in the surrounding sandy upland forests of the region, with over 12 years following restoration being primarily *Pinus taeda* and *Pinus palustris* on the drier locations (Richardson, unpublished data).

Despite a relatively uniform restoration strategy, both soil properties and vegetation are heterogenous across this wetland (Figures 19.3 and 19.4). This is primarily driven by the very gradual change in elevation and the corresponding changes in flooding parameters that occur across the site. Higher concentrations of soil C and phosphate correspond to both lower elevations and finer soil texture. Pattern in vegetation was associated primarily with soil types and elevation changes, but an additional purely spatial component may be related to dispersal (Hanchey 2002). Because this wetland is not homogeneous in its soil properties and vegetation, it is unlikely that all portions of the wetland are of equivalent function or that they will develop along identical trajectories. The lesson here is that anticipating soil heterogeneity as much as possible prior to restoration should aid in the proper selection of species to reintroduce and allow for more accurate predictions about the effects the restoration strategy will have at a site.

Created, Restored, and Natural Wetland Soil Comparisons

In contrast to natural wetlands (NWs), soils of created wetlands (CWs) and restored wetlands (RWs) thus appear to exhibit much lower spatial variability, but more recent and detailed studies show some notable exceptions among soil variables (Stolt et al. 2000; Bruland and Richardson 2005a, 2006). Several factors contribute to the homogeneity of CW/RW soils. First, wetland creation and restoration often involve the use of heavy machinery to remove topsoil and excavate into subsoil. In this process, soil surfaces are extensively cut and scraped, leaving flat, compacted surfaces with little relief (Clewell and Lea 1990; Stolt et al. 2000). Second, grading and site preparation activities tend to mix soils in both horizontal and vertical directions, disrupting soil zonation and horizons. Third, the use of uniform fill material or upland topsoil also leads to homogeneous soil conditions. Fourth, RWs are often located on former agricultural land. Long-term agricultural activity homogenizes the topsoil, which is most evident for attributes such as organic matter, nitrogen, and cation exchange capacity (Whisenant et al. 1995; Robertson et al. 1997; Gonzales et al. 2000). Over time, the combined action of physical, chemical, and biological processes can be expected to generate spatial heterogeneity in CW/RW soils comparable to that of NWs. However, these processes occur over time scales of decades to millennia rather than five-year jurisdictional monitoring periods (Mitsch and Wilson 1996; Craft et al. 2003; Hossler et al. 2011). Comparing the spatial distribution of soils in CW/RWs to paired NWs is important since it has been argued that environmental variability (which includes soils, topography, microclimate, etc.) and species richness are positively correlated (Williams 1964; Jeltsch et al. 1998; Ettema and Wardle 2002). If the soils of CW/RWs are more homogeneous than those of NWs, CW/RWs could be expected to also have lower diversity of soil biota and vegetation. Differences in the spatial characteristics of the soils of CWs/RWs and NWs may also indicate different controls on population and processes (Levin 1992), and lead to unsuccessful mitigation in the short-term.

To better understand patterns of spatial variability in soil properties of CWs, RWs, and NWs, Bruland and Richardson (2005a) sampled four CW/RW-NW pairs in the Coastal Plain of North Carolina using a spatially explicit design. Representative sites were selected from the following four different HGM subclasses: headwater riverine (stream order ≤2); mainstem riverine (stream order >2); non-riverine mineral soil flat; and non-riverine organic soil flat (see Chapter 16, Brinson 1993; Cole et al. 1997). The site pairs spanned a range of hydrogeomorphic (HGM) settings common in the Coastal Plain. In this study, the authors postulated that: (1) spatial variability of soil properties in riverine wetlands would be structured along gradients running perpendicular to streams, while spatial variability of soil properties in non-riverine wetlands would be structured in patches related to local factors (microtopography, vegetation); and (2) that soil properties of CW/RWs would exhibit less spatial variability than soil properties of NWs as prior land-use and mitigation activities tend to homogenize soil properties.

Trend surface analysis revealed that even in plots selected for homogeneous topography, linear and nonlinear trends were present in both CWs/RWs and NWs across all subclasses (Bruland and Richardson 2005a; Bruland et al. 2006, 2009). Further analysis indicated that fine-scale variability for Db, SOM, and pH was more prevalent in NWs than in paired CW/RWs. At certain sites, prior land-use and mitigation activities reduced spatial variability of soil properties such as sand content, while at other sites they increased or had no effect on variability of soil properties, such as SOM. Created/restored wetlands had higher mean Db, pH, and sand content, while NWs had higher SOM. Patterns of variability were complex and differed among soil properties, sites, and HGM subclasses. This lack of consistency suggested that spatial soil structure of wetlands was more than simply a function of wetland status (created-restored vs. natural) or hydrogeomorphoric setting (riverine vs. non-riverine) but rather was often related to the property or process being compared or presence of deep organic soils, which tended to make the natural sites more homogeneous than the restored sites for variables like SOM (Figure 19.5a). For example, SOM at Dismal Swamp (DS) was distributed homogeneously across the NW plot. Unlike the other three subclasses, which had mineral substrates, soils at DS were highly organic Histosols. As only a few woody and herbaceous species are adapted to live in such wet, acid, and organic soils, the litter that collects on the forest floor is fairly homogeneous. Furthermore, when surface organic horizons are thick, there is little if any mixing with underlying mineral horizons. This suggests that attempting to create heterogeneous distributions of SOM at non-riverine organic soil flats may not be appropriate. Moreover, these peatland sites may be the most difficult to create or restore, as peat formation is a process that occurs over periods of decades to millennia (Richardson 2008, 2012).

Interpolated maps of the soil properties among all the sites revealed homogeneous distributions of NO_3-N across the CW/RW plots compared to much more heterogeneous distributions of NO_3-N across the NW plots (Figure 19.5b). Regression analysis confirmed that either NO_3-N or soluble organic carbon were significant predictors of the denitrification enzyme activities (DEA) at each plot. Interpolated maps of predicted DEA generally showed similar patterns to those of NO_3-N (Figure 19.5b,c). While some nitrate and DEA hotspots were observed in the CW/RWs, more were present in the NWs. These results indicated that spatial distributions of soil chemical properties and DEAs were considerably different in CW/RWs than in NWs. This study also documented that CWs and RWs with homogeneous soil chemical distributions may not develop the full range of soil biogeochemical processes that occur in NWs. A better understanding of this phenomenon will help us to incorporate appropriate variability into wetland mitigation design and construction, improving creation, and restoration of functional wetlands.

As hypothesized, riverine sites displayed significant trends perpendicular to stream for soil properties such as Db, SOM, pH, and sand (Bruland and Richardson 2005b). Contrary to their hypothesis, soils of non-riverine wetland sites also displayed significant linear trends. Thus, it may be difficult to generalize about coarse-scale spatial trends in riverine versus non-riverine wetlands, as no consistent differences were observed. These results also hinted at the site-specific nature of spatial variability, in which unique geologic,

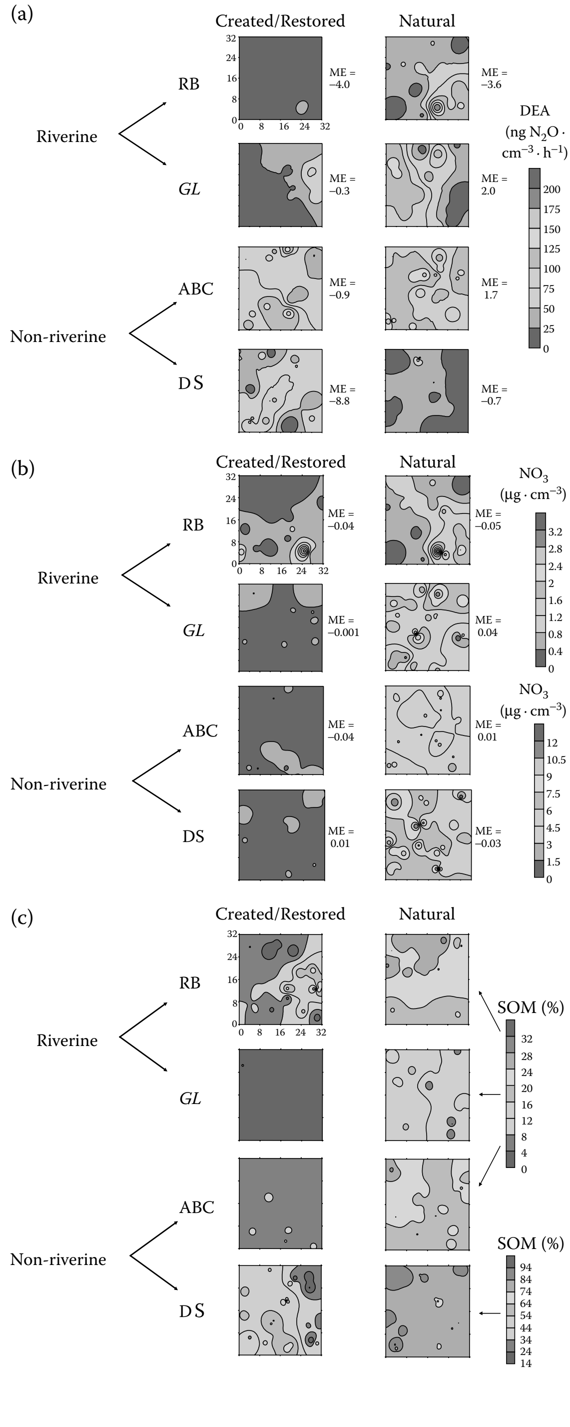


FIGURE 19.5
(See color insert.) (a) Spatial distribution of predicted denitrification enzyme activities (DEA), (b) spatial distribution of nitrate (NO_3-N), and (c) spatial distribution of soil organic matter (SOM) in restored and natural wetlands. Sites consist of Rowel Branch (RB), Grimesland (GL), ABC, and Dismal Swamp (DS). (Adapted from Bruland, G. L. and C. J. Richardson. 2006. *Wetl Ecol Manag* 14: 245–251; Bruland, G. L. and C. J. Richardson. 2005a. *Soil Sci Soc Am J* 69: 273–284.)

hydrologic, vegetative, and land-use histories may interact to create unique patterns of spatial variability. Further analysis indicated that prior land-use and mitigation activities might erase fine-scale spatial structure in CW/RW plots. This may occur as excavation, grading, and earth-moving activities mix soil patches horizontally and soil horizons vertically. Their analysis also indicated that land-use and mitigation activities can, in some cases, randomly create hotspots of soil characteristics rather than homogenize fine-scale spatial structure. Overall these results indicate that patterns of spatial variability of basic soils properties in CW/RWs and NWs are not consistent among variables as they appear to be influenced by a variety of factors including HGM setting, prior land-use, and mitigation activities. Fine-scale spatial randomness is most likely better for wetland function than fine-scale homogeneity, as randomly distributed soil properties would allow for more vegetative diversity and a wider range of edaphic conditions. Replacing the heterogeneous distributed soils of NWs with more homogeneously distributed soils of CW/RWs may not result in functional equivalency. The presence of spatial variability in soil properties from the CW/RWs may be due to the action of physical and biogeochemical processes or to the fact that prior land-use and mitigation activities may actually increase the variability of certain soil properties. A better understanding of these patterns will help us to incorporate variability into CW/RW design and to promote conditions that will allow for appropriate variability to develop, ultimately leading to improvements in creation and restoration of functional wetlands.

Soil Microtopography and Restoration

Microtopography is a characteristic feature of many NWs that is commonly lacking in RWs. Consequently, it has been suggested that microtopography must be reestablished in RWs to accelerate the development of wetland functions. Studies of natural freshwater wetlands have suggested that a key factor promoting vegetative structure and composition is microtopography (Barry et al. 1996; Vivian-Smith 1997). Causes of microtopographic heterogeneity in natural wetlands, although seldom documented, include sediment accumulation, erosion, tree fall, root growth, litterfall, animal burrowing, animal tracks, and variations in plant communities, which often create hummocks and tussocks. Scales of microtopographic variability in natural wetlands range from 0.01 m (as a result of sedimentation or animal tracks), to greater than one meter (following tree throw) (Vivian-Smith 1997). These features not only contribute to the vegetative structure of natural wetlands but also to their function (Ettema and Wardle 2002; Bruland and Richardson 2005a). For example, microtopography creates both aerobic and anaerobic zones that are needed for nitrogen retention and transformation (Reddy and Patrick 1976, 1984). Likewise, microtopographic variation in a floodplain forest in Georgia regulated Al and Fe oxide content and significantly affected biogeochemical cycling (Darke and Walbridge 2000). Thus, microtopography creates a mosaic of soil patches with substrates that differ structurally, hydrologically, and chemically (Bledsoe and Shear 2000).

It has been suggested that the re-creation of microtopography may improve restoration success (Barry et al. 1996; Cantelmo and Ehrenfeld 1999). The fact that microtopography has been reported to be missing in most restored wetlands (Barry et al. 1996; Whittecar and Daniels 1999; Stolt et al. 2000) is an unfortunate result of many restoration sites being located on formerly leveled agriculture lands and no effort to create a diversity of microhabitats (i.e., hummocks and hollows) in the soil is attempted due to time and cost constraints. Furthermore, minimal data exist on the length of time needed for restored wetlands to develop microtopography that is representative of natural wetlands.

In 2005, Bruland and Richardson undertook a study looking at the effects of recreating surface microtopography during wetland restoration or creation as an effective way to accelerate the development of wetland functions. The study addressed this proposition by investigating a restored wetland site that contained elevated hummocks (mounds) and lower elevation hollows (depressions), on otherwise level land (flats) of intermediate elevation in coastal North Carolina (Bruland and Richardson 2005b). The following hypotheses were tested: (1) denitrification will be greatest in the hollows due to the prevalence of inundated/saturated conditions in this zone compared to flats and hummocks; (2) plant species richness will be lower in the zones experiencing moisture extremes (wetter hollows and drier hummocks) than in the flats; and (3) aboveground biomass will be highest in hollows on account of the growth of *Typha* and *Scirpus* that prefer hollows to flats and hummocks.

The restoration site was a non-riverine mineral soil flat according to the HGM Classification System (Brinson 1993). The hummock, hollow, and flat plots in the study area were located in an area with soils from the Leaf series (fine, mixed, active, thermic Typic Albaquults). In the 1960s, much of this area was cleared, ditched, and converted into agricultural fields. The former agricultural field, of approximately 37 ha, was restored to wetland status in 2001. The first step consisted of removing topsoil and stockpiling it onsite. Next, impervious ditch plugs were installed, ditches were filled, and surface microtopography was recreated with hummocks and hollows. The hummocks, approximately 1 m in height and 1.5 m in diameter, were designed to mimic tip-up mounds that occur from tree fall in local reference wetlands. The hollows, with a maximum depth of approximately 0.3 m and an average surface area of 20–40 m^2, were designed to match depressions also observed in nearby natural wetlands (North Carolina Department of Transportation 1999). Finally, the topsoil was reapplied across the entire site including the hollows and hummocks.

The study showed that the mean water table depth in the hollows was 10 cm above the soil surface for 146 out of 216 days while the mean water table depth in the flats was never above 10 cm. Mean water table depth in the hollows was above the soil surface for 179 days or 83% of this period compared to 87 days or 40% of this period in the flats. Although the water table was consistently higher in the hollows than in the flats, the water table in both zones tended to follow the same pattern over the growing season and during individual storm events (Bruland and Richardson 2005b). There were no significant differences in Db ($p = 0.86$) (hummock = 1.30, flat = 1.27, hollow = 1.33 g/cm^3) or SOM across the microtopographic zones. Significant microtopography by time interactions for soil temperature ($p < 0.05$) and moisture ($p < 0.001$), was found indicating that differences between zones were not consistent throughout the growing season. Hummocks had significantly higher nitrate ($p < 0.0001$) and ammonium ($p = 0.001$) than flats and hollows for most of the growing season. Significant differences in microbial biomass carbon (MBC) and DEA across the microtopographic zones were not detected but the hollows had higher values (Figure 19.6a). However, plant species richness and biomass were significantly different ($p < 0.001$) across the microtopographic zones. Flats supported the greatest numbers of wetland species with hummocks < hollows < flats. Aboveground biomass across the microtopographic zones followed a different pattern than richness: hummocks < flats < hollows, owing to the growth of emergent wetland herbs in hollows.

Thus, microtopographic re-establishment had significant effects on vegetative diversity and productivity. In fact, effects on vegetation were possibly more pronounced than on hydrology and soil properties. For example, by the end of the third growing season since restoration, species richness was, as hypothesized, significantly higher in the flats than in the hummocks or hollows. In the flats, intermediate hydrologic and edaphic conditions allowed upland, facultative wetland, and obligate wetland species to compete with planted wetland tree seedlings for nutrients and light. The higher elevation of the hummocks made them favorable for plants species that are less tolerant of flooding and require a more mineral soil for survival (Titus 1990). The saturated and inundated conditions in the hollows, on the other hand, were tolerable only to a few obligate emergent marsh species that grew rapidly in these areas. There was little overlap of species among the different microtopographic zones, with most species preferring a certain microtopographic zone. Such interspecific differences in habitat preferences during vegetative establishment have been observed previously in a drained Wisconsin marsh (Zedler and Zedler 1969), across a water level gradient (Keddy and Ellis 1984), and in experimental wetland mesocosms with artificial hummocks and hollows (Vivian-Smith 1997).

As postulated, hollows had significantly higher biomass than flats or hummocks due to the rapid growth of emergent marsh species such as *S. cyperinus*, *J. effusus*, and *T. latifolia* in these areas (Figure 19.6b). The more diverse species assemblage in the flats produced significantly less aboveground biomass than the hollows possibly due to competitive interactions (i.e., shading, root uptake) among the different species. Likewise, in the higher and drier hummocks, moisture limitation and intense insect and faunal activity may have limited production of aboveground biomass. Had the entire site been graded to the low elevations found in the hollows, hydrologic conditions may have been too wet for a forested wetland and, instead, an emergent freshwater marsh may have developed. On the other hand, if the entire site had been graded to the elevation of the flats, the dense herbaceous vegetation and pockets of open water probably would not have developed. These results were similar to previous research that reported individual plant species (Eldridge et al. 1991) and entire plant communities to be related to microtopographic heterogeneity (Ehrenfeld 1995).

The reestablished microtopography at the restoration site provided a variety of hydrologic, edaphic, and vegetative conditions at any given time over the course of the growing season. If hydrologic conditions change in the future, as they undoubtedly will with the development of the vegetation and closing of the forest canopy, there is a greater chance that a site with reestablished microtopography will continue to provide additional suitable hydrologic conditions than a site of uniform microtopography (Barry et al. 1996). While it may be too costly to reestablish microtopography across large created or restored sites in their entirety, there is value in reestablishing microtopography on certain sections or subplots. Wetland designers and engineers should be encouraged and allowed to develop restoration plans that contain hummocks and hollows that are consistent with the microtopography of nearby natural wetlands of the same hydrogeomorphic setting. As with Tweedy and Evans (2001) and Cantelmo and Ehrenfeld (1999), results from this study clearly show that microtopographic reestablishment, especially when sites are located on smooth, flat, former agricultural land, will result in reestablishment of numerous hydrologic, edaphic, and vegetative benefits.

Soil Carbon and Restoration

Although we note the difference among authors in using the terms soil organic matter (SOM), organic matter (OM), and compost, here we simply use them interchangeably to simply represent the carbon (C) content or soil organic carbon (SOC) additions to the soil in the following studies. When compared to NWs, soils of CWs usually have higher Db and lower levels of OM (Clewell and Lea 1990; Bishel–Machung et al. 1996; Shaffer and Ernst 1999; Stolt et al. 2000; Bruland and Richardson 2006). The fact that litter layers in CWs are often poorly developed or absent further confounds the problem (Hunter and Faulkner

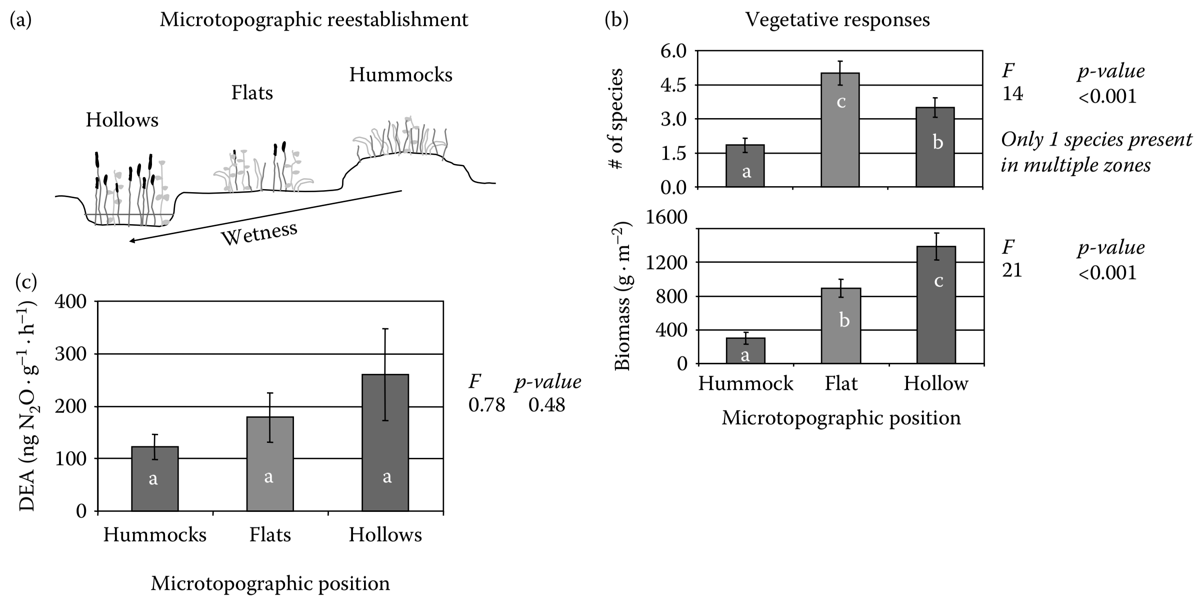


FIGURE 19.6
(a) Illustrated representation of the microtopographic features reestablished at the wetland restoration site. (b) Mean aboveground standing biomass and species richness for the hummock, flat, and hollow plots sampled. Bars represent means ± 1 SE, and means with different letters are significantly different, and (c) mean aboveground denitrification enzyme activity (DEA) for the hummock, flat, and hollow plots sampled. Bars represent means ± 1 SE, and means with different letters are significantly different. (Adapted from Bruland, G. L. and C. J. Richardson. 2005b. *Restor Ecol* 13(3): 1–9.)

2001). Low OM levels have been shown to limit plant establishment and growth (Zedler and Langis 1991; Stauffer and Brooks 1997; van der Valk et al. 1999) as well as nutrient cycling in restored wetlands (Groffman et al. 1996). Wetland disturbance often involves ditching and/or draining the wetland which creates a more aerobic soil environment in which SOM is oxidized and soil storage of C decreases (Schlesinger and Bernhardt 2013). Some restorations involve excavations to intersect ground water and can expose coarse subsoil (Stauffer and Brooks 1997; Sutton-Grier et al. 2009). Soils are also disturbed and compacted during restoration due to heavy equipment at the project site, which is necessary for re-grading the site to restore wetland hydrology (Unghire et al. 2011). This disturbance and compaction changes the structure of the soil and increases the Db making it more difficult for plant roots to penetrate soils (Clewell and Lea 1990). Re-grading may also involve the removal of the topsoil layers, which tend to be richest in SOM and microbial populations. Thus, it is perhaps not surprising that CWs/RWs tend to lack SOM and microbes in comparison to NWs, especially in the early years of restoration (Ballantine and Schneider 2009).

Soil properties, especially SOM, have been shown to be one of the slowest ecosystem components to develop after restoration or creation (Craft et al. 2002, 2003; Ballantine and Schneider 2009; Card et al. 2010) and may in some cases be impossible to restore due to other constraints on the ecosystem (Zedler and Langis 1991; Simenstad and Thom 1996; Zedler and Kercher 2005). Because restored wetland ecosystems may need many years to develop conditions that match those of their natural counterparts, ecosystem functioning in restored sites may be limited (Shaffer and Ernst 1999). A case in point being the study by Ballantine and Schneider (2009) where they showed that even in depressional wetlands the top 5 cm of soil, SOM, Db, and cation exchange capacity (CEC) achieved only 50% of reference levels 55 years after restoration. Soil development processes in these depressional wetlands appeared to be driven by autochthonous inputs and by internal processes such as litter decomposition and were not accelerated in the initial phase of development by allochthonous inputs as has been documented in coastal salt marshes and riverine floodplains (Craft et al. 2003; Hogan et al. 2004). Also recent models have suggested that it might take up to 300 years for a newly created wetland to sequester the amount of SOC found in an NW (Hossler and Bouchard 2010). Therefore, it would be very beneficial to be able to mitigate C limiting conditions at the start of a restoration to "jump-start" restored wetlands so that they could more quickly attain levels of ecosystem function more similar to their natural counterparts (Sutton-Grier et al. 2009). As SOC is an important determinant of wetland functions such as nitrate reduction, the transformation of organic nutrients to inorganic plant-available forms as well as helping maintain soil moisture and supporting plant growth, adding or maintaining SOC is considered essential to maintaining or restoring key ecosystem functions and habitat.

Because SOM is critical to wetland ecosystem functioning, at sites where soils have been disturbed or removed, restoring SOM may be critical to effective restoration. This is a major problem in SOM-deficient sites due to the slow rate of SOM buildup, which occurs naturally on decadal to millennial scales. This suggests that during the selection of restoration sites, one should first consider sites with adequate SOM already present in sufficient amounts to maintain soil moisture and support plant growth. When this is not possible then the question remains as to what is the most effective method to restore SOM. One suggested method of soil conditioning is to spread and disk organic matter into soils, particularly composted sludge because it introduces soil microfauna and can improve soil fertility (Clewell and Lea 1990). Topsoil additions have been shown to increase SOM content, as well as soil moisture, water holding capacity, and P sorption (Bruland and Richardson 2004; Bruland et al. 2009). There are a variety of suitable organic wastes that could potentially serve as SOM amendments including municipal leaf and lawn compost, certain sewage sludge/biosolids materials, food processing waste, and forest products (Stauffer and Brooks 1997). Instead of occupying valuable space in municipal landfills, these types of organic wastes could be used to improve soil conditions in CWs. For example, SOM amendments applied to a CW in Pennsylvania (Stauffer and Brooks 1997) significantly increased soil moisture and nitrate-N availability compared to unamended control plots. Stauffer and Brooks (1997) concluded that amendments should be considered if created wetland projects contain <10% SOM. Another study of a CW in Massachusetts reported that OM amendments produced high levels of microbial activity (Duncan and Groffman 1994). However, another study of SOM amendments at a created salt marsh in southern California revealed that soil C and N pools were not increased by amendments due to high decomposition rates in that site's sandy soils (Gibson et al. 1994). Furthermore, the studies in Pennsylvania and Massachusetts only added a single level of OM. As amendments are expensive, especially when applied to larger sites, it is important to review studies that try and determine optimal amendment levels to maintain ecosystem functions. The importance of adding SOM to aid in the "jump starting" of restoration wetland functions is demonstrated by highlighting key findings and recommendations in recent studies by Bruland et al. (2006), Sutton-Grier et al. (2009), and others.

An urban wetland/stream restoration located in Charlotte, NC project included remeandering the stream and establishing two types of adjacent wetlands, deeper (wetter) "low marsh" areas, and shallower (drier) "high marsh" areas. The soils at the system were originally Monacan loam (a sandy loam), but when homes were built on the system, 2 m of fill material was added to the system. As part of the restoration, this fill was removed resulting in very disturbed and SOM- and nutrient-poor soils. Initially, topsoil and different amounts of organic material (a combination of topsoil, wood chips, and pathogen-free biosolids from wastewater plants) were added to the system. After compost incorporation, vegetation was planted in the floodplain including *Peltandra virginica* (arrow arum), *Pontederia cordata* (pickerelweed), *Sagittaria latifolia* (duck potato), and *Schoenoplectus tabernaemontani* (softstem bulrush).

Results showed available N and P increased with increasing soil organic matter in both the low and high marsh (Sutton-Grier et al. 2009). Total microbial biomass (MB) and microbial activity (measured by denitrification potential (DEA, Figure 19.7)) also significantly increased with increased OM in both marsh communities, as did soil moisture. Thus, compost amendments were an effective method for increasing soil properties, positively influencing soil available N, P, microbial biomass, and moisture as well as ecosystem functions including nutrient cycling (i.e., DEA), but had limited early impacts on plant communities' richness and growth (Sutton-Grier et al. 2009).

A longer-term study site was located in the Virginia coastal plain physiographic region (Bruland et al. 2009). This site was part of the Virginia DOT compensatory mitigation program used primarily to offset non-tidal forested wetland impacts. The pre-existing soil series at the site were a complex of Chickahominy (fine, mixed, semiactive, thermic Typic Endoaquults) and Newflat (fine, mixed, subactive, thermic Aeric Endoaquults) soils (Bergschneider 2005). The site was originally an upland mixed hardwood forest that had been partially converted into an agricultural field. Mitigation efforts attempted to convert the fields and remaining forest remnants into wetland status by removing O + A + E horizons and excavating into the subsoil (Btg horizon) to an elevation presumed to be indicative of the seasonal high water table. However, due to soil compaction, high silt and clay content, and lack of SOM in the subsoil, vegetation that established over certain portions of the site was dominated by facultative upland or obligate upland species. This indicated

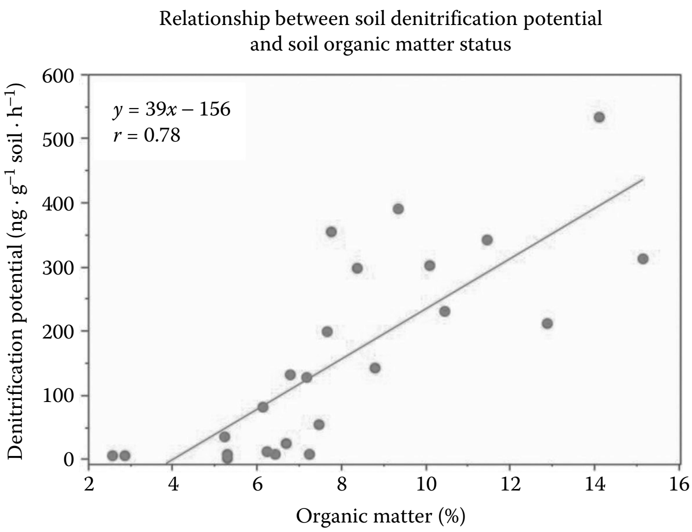


FIGURE 19.7
Denitrification potential versus organic matter in soils following compost additions in high and low marshes at a restoration site in Charlotte, North Carolina. (Adapted from Sutton-Grier, A. E., M. Ho, and C. J. Richardson. 2009. *Wetlands* 29(1): 343–352.)

that the hydrologic and edaphic conditions in certain parts of the site were not appropriate for supporting development of wetland vegetation. For example, while the wetter zone clearly supported wetland vegetation, there were areas of the drier zone that would not have met the vegetative success criteria.

To improve vegetative growth and survival, five different levels of compost were added in to the wetter and drier zones of the site (Bergschneider 2005). The two zones were differentiated by both elevation and vegetation, with the wetter zone being in slightly lower topographic position and supporting a plant community of more emergent marsh species such as *Typha* spp. and *Scirpus* spp. than the drier zone which contained a significant component of *Lespedeza cuneata*. Stable wood-fines compost was applied to replicate plots (4x) in each zone at the following levels: 0, 56, 112, 224, and 336 Mg ha^{-1} after pre-existing vegetation was mowed and the surface was root-raked. The compost was incorporated with an offset disk followed by a rototiller into the upper 10 cm of soil (Bruland et al. 2009).

The study revealed several significant and interesting effects of adding SOM in terms of edaphic properties and ecosystem responses (Bruland et al. 2009). Specifically, Db decreased exponentially with increasing amendment level (Figure 19.8a). Soil pH displayed a linear decrease with amendment level in both wetter and drier zones (Figure 19.8b); soluble organic carbon exhibited linear increases with amendment level across both wetter and drier zones (Figure 19.8c). While not statistically significant, the model for MBC in the drier zone showed a trend of increasing MBC with amendment level (Figure 19.8d). In the wetter zone, MBC showed a significant linear relationship with amendment level. The DEA exhibited a nonlinear relationship to amendment level, and the drier zone 112 Mg ha^{-1} plots and the wetter zone 224 Mg ha^{-1} plots had higher DEA rates than the

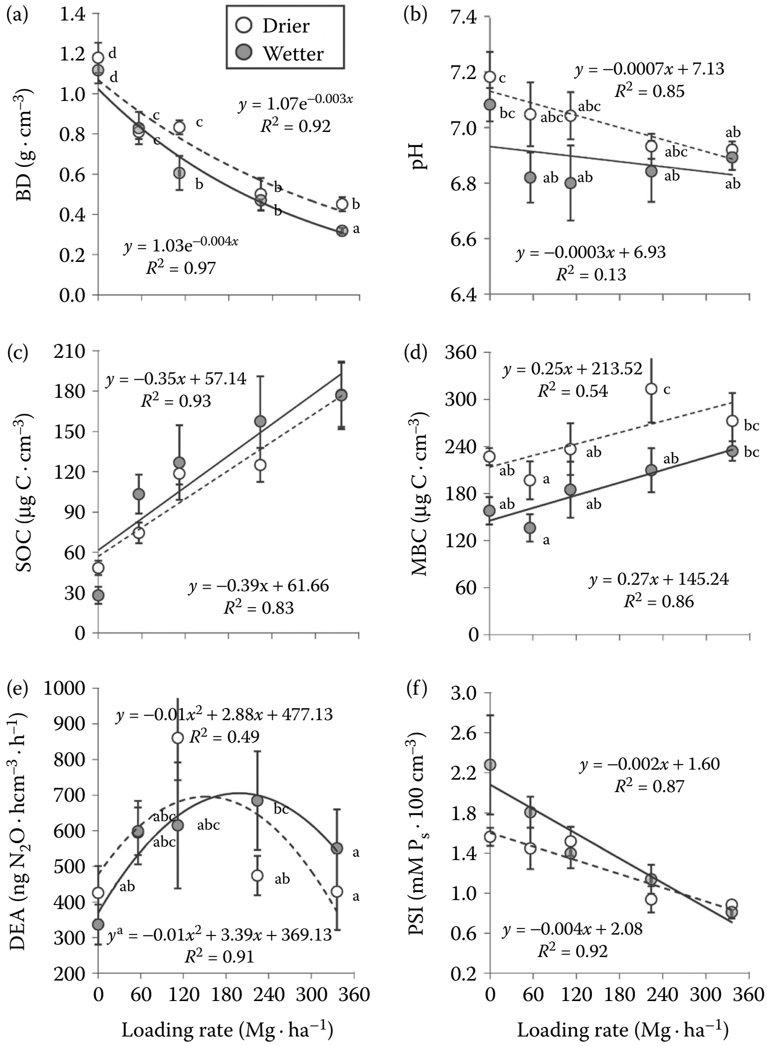


FIGURE 19.8
Regressions of (a) bulk density (Db), (b) pH, (c) soluble organic carbon (SOC), (d) microbial biomass carbon (MBC), (e) denitrification enzyme assay (DEA), and (f) the phosphorus sorption index (PSI) as a function of organic amendment loading rate. Circles represent the mean value at each loading rate and error bars represent ± 1 standard error. Open circles represent data from the drier zone while shaded circles represent data from the wetter zone. For each panel, the upper equation and R^2 value correspond to plots from the drier zone and the lower equation and R^2 correspond to plots from the wetter zone. Circles with different letters are significantly different according to the ANOVA post-hoc LSD test. (From Bruland, G. L. et al. 2009. *Wetlands* 29(4): 1153–1165.)

control plots (Figure 19.8e). The two treatments with the highest mean DEA rates were the 112 Mg ha^{-1} drier plots and the 224 Mg ha^{-1} wetter plots. The phosphorus sorption index (PSI) decreased with SOM additions (Figure 19.8f).

Findings from this study suggest that the optimal organic amendment level for mitigation wetlands in the coastal plain of Virginia is between 60 and 180 Mg ha^{-1} as this range of additions appeared to give the optimal reductions in Db, and increases MBC and DEA potential, without resulting in detrimental decreases in pH or P sorption. Thus, the 60–180 Mg ha^{-1} range appeared to balance economic constraints with nutrient transformation and retention processes to provide the created wetland with maximum functional benefits. In studying the response of the vegetation to the organic amendment loading rates at this same site, Bailey et al. (2007) concluded that the amendment loading rate of 112 Mg ha^{-1} was optimal as it provided soil nutrient levels similar to natural wetlands in this region as well as minimized changes in the soil surface elevation due to the added amendment material. We have also provided some more general recommendations about organic amendments and microtopographic reestablishment in restored wetlands (Table 19.1).

While it may be too costly to amend large mitigation sites in their entirety, this study shows that there may be value in amending certain sections or subplots of sites. Ultimately, just as there are hydrologic and vegetative success criteria for created wetland, there should be edaphic success criteria as well. When poor soil conditions can lead to inadequate hydrology and low plant survival, establishing proper substrate conditions may be as important as reestablishing wetland hydrology and hydrophytic vegetation. With the development of edaphic success criteria, sites that did not meet an SOM threshold would need to be amended during mitigation to improve functionality. Such OM thresholds are also now a standard recommendation in Virginia (Daniels et al. 2005).

Together, these studies suggest that there are several benefits of SOM additions to soil nutrient supplies, microbial communities, and nutrient cycling. Based on the findings of Bruland et al. (2009) and Sutton-Grier et al. (2009) there are, however, a few factors to consider before deciding to apply SOM amendments to a restoration site. First, it is important to recognize that applying compost amendments is a fairly labor-intensive process that involves both spreading the compost and mixing it into the topsoil horizon. Second, it can be costly to obtain sufficient compost for an entire restoration site. Also if the SOM is not well-mixed into the top soil horizon, some of the compost can be washed away in the first storm event or transported to other parts of the site. Thus, adding compost may be an effective way to increase SOM levels, but it may not be a feasible solution to implement at all restoration sites because it involves considerable time, effort, and cost. In addition, other research has suggested that the type of SOM as well as the amount of SOM applied may impact how SOM additions influence microbial populations and activities (Saison et al. 2006; Lou et al. 2007). Lou et al. (2007) determined that rice straw contained more organic C and N and less lignin and was therefore a higher quality compost input that enhanced microbial activity more than addition of rice roots. Saison et al. (2006) determined that

TABLE 19.1

General Recommendations for Organic Matter and Microtopographic Reestablishment in Restored Wetlands

	Organic Amendments	Microtopographic Reestablishments
Wetland Restoration Guidelines	Efforts should be made to match the soil organic matter (SOM) of reference wetlands in similar hydrogeomorphic subclasses of the region where the restoration occurs. If no such information is available, then SOM amendments should be considered if SOM levels at the restoration site are <2.5%.	Efforts should be made to match the spatial pattern of microtopography (i.e., hummocks and hollows, or tussocks, etc.) of reference wetlands in similar HGM subclasses in the region where the restoration occurs. This is especially important at former agricultural sites or other sites where human land-use created artificially flat microtopographic conditions.

there were important shifts in microbial community composition when high levels of SOM were added, but they observed little to no change in community composition with low levels of SOM addition. Future work should focus on how different amounts and types of SOM affect ecosystem properties and functions in restored wetlands.

Conclusion

Collectively, these studies show that wetland restoration projects must consider edaphic conditions in conjunction with hydrologic and vegetation plans in order to optimize and "jump-start" ecosystem structure and function. Specifically, projects should consider the spatial variability needed in the created or restored wetlands to mimic the natural conditions for each wetland type. Planting success will depend on this knowledge. The influence of microtopography and soil structure on processes like denitrification and plant community structure validates their importance in creating diverse wetland habitats and enhancing biogeochemical processes. The organic amendment studies suggest that restoration projects, especially those involving highly disturbed soils, should include SOM in the project design to promote soil ecosystem functions, particularly nutrient cycling. Without supplemental SOM, the soil properties of many restored wetlands may never be equivalent to those of natural wetlands for decades, limiting the effectiveness of restoration. Just as there are hydrologic and vegetative success criteria for restored and created wetlands that are included in the regulatory mitigation process, the studies described in this review make a strong case that attempts should be made to incorporate edaphic success criteria into the jurisdictional wetland mitigation process.

Acknowledgments

Thanks go to the many students and lab technicians who worked on the main Duke wetland Center projects reviewed in this chapter. The methods and key findings of these studies are found in detail in the articles cited in this review. Randy Neighbarger helped with the formatting and editing. Funding for studies came from the Duke Wetland Center Endowment, City of Charlotte, the Peterson Foundation, and Clean Water Management Trust Fund of NC.

References

Bailey, D. A., J. E. Perry, and W. L. Daniels. 2007. Vegetation dynamics in response to organic matter loading rates in a created freshwater wetland in southeastern Virginia. *Wetlands* 27: 936–950.

Balcombe, C. K., J. T. Anderson, R. H. Fortney, J. S. Rentch, W. N. Grafton, and W. S. Kordek. 2005. A comparison of wetland plant communities in mitigation and reference wetlands in the Mid-Appalachians. *Wetlands* 25: 130–142.

Ballantine, K. and R. Schneider. 2009. Fifty-five years of soil development in restored freshwater depressional wetlands. *Ecol Appl* 19(6): 1467–1480.

Barry, W. J., A. S. Garlo, and C. A. Wood. 1996. Duplicating the mound-and-pool microtopography of forested wetlands. *Restor Manage Notes* 14: 15–21.

Bergschneider, C. R. 2005. Determining an appropriate organic matter loading rate for a created coastal plain forested wetland. M.S. thesis. Virginia Tech, Blacksburg, VA, USA. http://www.lib.vt.edu/find/etds.php

Bishel-Machung, L., R. P. Brooks, S. S. Yates, and K. L. Hoover. 1996. Soil properties of reference wetlands and wetland creation projects in Pennsylvania. *Wetlands* 16: 532–541.

Bledsoe, B. P. and T. H. Shear. 2000. Vegetation along hydrologic and edaphic gradients in a North Carolina Coastal Plain creek bottom and implications for restoration. *Wetlands* 20: 126–147.

Bridgham, S. D., C. A. Johnston, and J. P. Schubauer-Berigan. 2001. P sorption dynamics in soils and coupling with surface and pore water in riverine wetlands. *Soil Sci Soc Am J* 65: 577–588.

Brinson, M. M. 1993. Changes in the functioning of wetlands along environmental gradients. *Wetlands* 13: 65–74.

Brinson, M. M. and R. Rheinhardt. 1996. The role of reference wetlands in functional assessment and mitigation. *Ecol Appl* 6: 69–76.

Bruland, G. L., M. F. Hanchey, and C. J. Richardson. 2003. Effects of agriculture and wetland restoration on hydrology, soils, and water quality of a Carolina Bay complex. *Wetl Ecol Manag* 11: 141–156.

Bruland, G. L. and C. J. Richardson. 2004. Hydrologic gradients and topsoil additions affect soil properties of Virginia created wetlands. *Soil Sci Soc Am J* 68: 2069–2077.

Bruland, G. L. and C. J. Richardson. 2005a. Spatial variability of soil properties in created, restored, and paired natural wetlands. *Soil Sci Soc Am J* 69: 273–284.

Bruland, G. L. and C. J. Richardson. 2005b. Hydrologic, edaphic, and vegetative responses to Microtopographic reestablishment in a restored wetland. *Restor Ecol* 13(3): 1–9.

Bruland, G. L. and C. J. Richardson. 2006. Comparison of soil organic matter in created, restored and paired natural wetlands in North Carolina. *Wetl Ecol Manag* 14: 245–251.

Bruland, G. L. C. J. Richardson, and S. C. Whalen. 2006. Spatial variability of denitrification potential and related soil properties in created, restored, and paired natural wetlands. *Wetlands* 24(4): 1042–1056.

Bruland, G. L., C. J. Richardson, and W. L. Daniels. 2009. Microbial and geochemical responses to organic matter amendments in a created wetland. *Wetlands* 29(4): 1153–1165.

Cantelmo, A. J. and J. G. Ehrenfeld. 1999. Effects of microtopography on mycorrhizal infection in Atlantic white cedar (*Chamaecyparis thyoides* (L.) Mills.). *Mycorrhiza* 8: 175–180.

Card, S. M., S. A. Quideau, and S. W. Oh. 2010. Carbon characteristics in restored and reference riparian soils. *Soil Sci Soc Am J* 74: 1834–1843.

Casanova, M. T. and M. A. Brock. 2000. How do depth, duration and frequency of flooding influence the establishment of wetland plant communities? *Plant Ecol* 147: 237–250.

Clewell, A. F. and R. Lea. 1990. Creation and restoration of forested wetland vegetation in the southeastern United States. In J. A. Kusler, and M. A. Kentula (Eds.) *Wetland Creation and Restoration: The Status of the Science*. Island Press, Washington, DC, pp. 195–232.

Cole, C. A., R. P. Brooks, and D. H. Wardrop. 1997. Wetland hydrology as a function of hydrogeomorphic subclass. *Wetlands* 17: 456–467.

Craft, C., S. Broome, and C. Campbell. 2002. Fifteen years of vegetation and soil development after brackish-water marsh creation. *Restor Ecol* 10: 248–258.

Craft, C., P. Megonigal, S. Broome et al. 2003. The pace of ecosystem development of constructed *Spartina alterniflora* marshes. *Ecol Appl* 13: 1417–1432.

Craft, C. B., S. W. Broome, and E. D. Seneca. 1988. Nitrogen, phosphorus and organic-carbon pools in natural and transplanted marsh soils. *Estuaries* 11: 272–280.

Daniels, W. L., G. Fajardo, C. R. Bergschneider et al. 2005. *Effects of Soil Amendments and Other Practices Upon the Success of the Virginia Department of Transportation's Non-tidal Wetland Mitigation Efforts*. Virginia Department of Transportation, Virginia Transportation Research Council, Charlottesville, VA, USA, VTRC05-CR25.

Darke, A. K. and M. R. Walbridge. 2000. Al and Fe biogeochemistry in a floodplain forest: Implications for P retention. *Biogeochemistry* 51: 1–32.
Duncan, C. P. and P. M. Groffman. 1994. Comparing microbial parameters in natural and constructed wetlands. *J Environ Qual* 23: 298–305.
Ehrenfeld, J. G. 1995. Microsite differences in surface substrate characteristics in *Chamaecyparis* swamps of the New Jersey pinelands. *Wetlands* 15: 183–189.
Eldridge, D. J., M. Westoby, and K. G. Holbrook. 1991. Soil-surface characteristics, microtopograohy and proximity to mature shrubs: Effects on survival of several cohorts of *Atriplex vesicaria* seedlings. *J Ecol* 78: 357–364.
Ettema, C. E. and D. A. Wardle. 2002. Spatial soil ecology. *Trends Ecol Evol* 17: 177–183.
Fennessy, M. S. and W. J. Mitsch. 2001. Effects of hydrology on spatial patterns of soil development in created riparian wetlands. *Wetlands Ecol Manag* 9: 103–120.
Gibson, K. D., J. B. Zedler, and R. Langis. 1994. Limited response of cordgrass (*Spartina foliosa*) to soil amendments in a constructed marsh. *Ecol Appl* 4: 757–767.
Goovaerts, P. 1998. Geostatistical tools for characterizing the spatial variability of microbiological and physico-chemical soil properties. *Biol Fert Soils* 27: 315–334.
Grace, J. B., L. Allain, and C. Allen. 2000. Vegetation associations in a rare community type—Coastal tallgrass prairie. *Plant Ecol* 147: 105–115.
Groffman, P. M., G. C. Hanson, E. Kiviat, and G. Stevens. 1996. Variation in microbial biomass and activity in four different wetland types. *Soil Sci Soc Am J* 60: 622–629.
Hanchey, M. F. 2002. *Heterogeneity in Soil and Vegetation Properties of a Restored Carolina Bay Wetland*. M.S. thesis. Duke University, Durham, NC, USA.
Hassink, J. 1995. Decomposition rate constants of size and density fractions of soil organic-matter. *Soil Sci Soc Am J* 59: 1631–1635.
Hogan, D. M., T. E. Jordan, and M. R. Walbridge. 2004. Phosphorus retention and soil organic carbon in restored and natural freshwater wetlands. *Wetlands* 24: 573–585.
Hossler, K. and V. Bouchard. 2010. Soil development and establishment of carbon-based properties in created freshwater marshes. *Ecol Appl* 20: 539–553.
Hossler, K., V. Bouchard, M. S. Fennessy, S. D. Frey, E. Anemaet, and E. Herbert. 2011. No-net-loss not met for nutrient function in freshwater marshes: Recommendations for wetland mitigation policies. *Ecosphere* 2(7): http://www.esajournals.org/doi/abs/10.1890/ES11-0009.1
Hunter, R. G. and S. P. Faulkner. 2001. Denitrification potential in restored and natural bottomland hardwood wetlands. *Soil Sci Soc Am J* 65: 1865–1872.
Jeltsch, F., S. J. Milton, W. R. J. Dean, N. Van Rooyen, and K. A. Moloney. 1998. Modeling the impact of small-scale heterogeneities on tree-grass coexistence in semi-arid ecosystems. *J Ecol* 86: 780–793.
Johnston, C. A., S. D. Bridgham, and J. P. Schubauer-Berigan. 2001. Nutrient dynamics in relation to geomorphology of riverine wetlands. *Soil Sci Soc Am J* 65: 557–577.
Keddy, P. A. and T. H. Ellis. 1984. Seedling recruitment of 11 plant species along a water level gradient: Shared or distinct response? *Can J Bot* 63: 1876–1879.
Kentula, M. E. 2000. Perspectives on setting success criteria for wetland restoration. *Ecol Eng* 15: 199–209.
Kirkman, L. K., P. C. Goebel, L. West, M. B. Drew, and B. J. Palik. 2000. Depressional wetland vegetation types: A question of plant community development. *Wetlands* 20: 373–385.
Lenssen, J. P. M., F. B. J. Menting, W. H. Van Der Putten, and C. W. P. M. Blom. 1999. Effects of sediment type and water level on biomass production of wetland plant species. *Aquat Bot* 64: 151–165.
Levin, S. A. 1992. The problem of pattern and scale in ecology. *Ecology* 73: 1943–1967.
Lou, Y. S., L. X. Ren, Z. P. Li, T. L. Zhang, and K. Inubushi. 2007. Effect of rice residues on carbon dioxide and nitrous oxide emissions from a paddy soil of subtropical China. *Water Air Soil Pollut* 178: 157–168.
Lyons, J. B., J. H. Gorres, and J. A. Amador. 1998. Spatial and temporal variability of phosphorus retention in a riparian forest soil. *J Environ Qual* 27: 895–903.
Mitsch, W. J. and J. G. Gosselink. 2007. *Wetlands*, 4th ed. John Wiley and Sons, Hoboken.

Mitsch, W. J. and R. F. Wilson. 1996. Improving the success of wetland creation and restoration with know-how, time, and self-design. *Ecol Appl* 6: 77–83.

NCDOT (North Carolina Department of Transportation). 1999. *ABC Wetland Mitigation Plan*. Office of Natural Environment and Roadside Environmental Unit, North Carolina Department of Transportation, Raleigh. N. C.

Osland, M. J., E. González, and C. J. Richardson. 2011. Coastal freshwater wetland plant community response to seasonal drought and flooding in northwestern Costa Rica. *Wetlands* 31: 641–652.

Pachepsky, Y. A., D. J. Timlin, and W. J. Rawls. 2001. Soil water retention as related to topographic variables. *Soil Sci Soc Am J* 65: 1787–1795.

Paz-Gonzalez, A., S. R. Vieira, and M. T. T. Castro. 2000. The effect of cultivation on the spatial variability of selected properties of an Umbric Horizon. *Geoderma* 97: 273–292.

Pezeshki, S. R., R. D. Delaune, and P. H. Anderson. 1999. Effect of flooding on elemental uptake and biomass allocation in seedlings of three bottomland tree species. *J Plant Nutr* 22: 1481–1494.

Poiani, K. A. and P. M. Dixon. 1995. Seed banks of Carolina bays—Potential contributions from surrounding landscape vegetation. *Am Midland Naturalist* 134: 140–154.

Reddy, K. R. and R. D. DeLaune. 2008. *Biogeochemistry of Wetlands*. Boca Raton: CRC Press.

Reddy, K. R. and W. H. Patrick. 1976. Effect of alternate aerobic and anaerobic conditions on redox potential, organic matter decomposition and nitrogen loss in a flooded soil. *Soil Biol Biochem* 8: 491–495.

Reddy, K. R. and W. H. Patrick. 1984. Nitrogen transformations and loss in flooded soils and sediments. *CRC Critic Rev Environ Contamin* 13: 273–309.

Reese, R. E. and K. K. Moorhead. 1996. Spatial characteristics of soil properties along an elevational gradient in a Carolina bay wetland. *Soil Sci Soc Am J* 60: 1273–1277.

Richardson, C. J. 2008. *The Everglades Experiments: Lessons for Ecosystem Restoration*. Springer-Verlag, New York.

Richardson, C. J. 2012. Pocosins: Evergreen shrub bogs of the Southeast. In D. P. Batzer, and A. H. Baldwin (Eds.) *Wetland Habitats of North America: Ecology and Conservation Concerns*. University of California Press, Berkeley, pp. 189–202.

Richardson, C. J., N. Flanagan, M. Ho, and J. Pahl. 2011. Integrated stream and wetland restoration: A watershed approach to improved water quality on the landscape. *Ecol Eng* 37: 25–39.

Richardson, C. J. and J. W. Gibbons. 1993. Pocosins, Carolina bays, and mountain bogs. In W. H. Martin, S. G. Boyce, and A. C. Echternacht (Eds.) *Biodiversity of the Southeastern United States*. John Wiley & Sons, Inc., New York, USA, pp. 257–309.

Robertson, G. P., K. M. Klingensmith, M. J. Klug, E. A. Paul, J. R. Crum, and B. G. Ellis. 1997. Soil resources, microbial activity, and primary production across an agricultural ecosystem. *Ecol Appl* 7: 158–170.

Saison, C., V. Degrange, R. Oliver et al. 2006. Alteration and resilience of the soil microbial community following compost amendment: Effects of compost level and compost-borne microbial community. *Environ Microbio* 8: 247–257.

Schlesinger, W. H. and E. Bernhardt. 2013. *Biogeochemistry: An Analysis of Global Change*. Academic Press, Amsterdam.

Seabloom, E. W., K. A. Moloney, and A. G. Van Der Valk. 2001. Constraints on the establishment of plants along a fluctuating water-depth gradient. *Ecology* 82: 2216–2232.

Shaffer, P. W. and T. L. Ernst. 1999. Distribution of soil organic matter in freshwater emergent/open water wetlands in the Portland, Oregon Metropolitan Area. *Wetlands* 19: 505–516.

Sharitz, R. R. and J. W. Gibbons. 1982. *The Ecology of Southeastern Shrub Bogs (Pocosins) and Carolina Bays a Community Profile*. U.S. Fish and Wildlife Service, Division of Biological Services, Washington, DC, USA. FWS/OBS-82/04.

Simenstad, C. A. and R. M. Thom. 1996. Functional equivalency trajectories of the restored Gog-Le-Hi-Teestuarine wetland. *Ecol Appl* 6: 38–56.

Stauffer, A. L. and R. P. Brooks. 1997. Plant and soil responses to salvaged marsh surface and organic matter amendments at a created wetland in Pennsylvania. *Wetlands* 17: 90–105.

Stolt, M. H., M. H. Genthner, W. L. Daniels, and V. A. Groover. 2001. Spatial variability in palustrine wetlands. *Soil Sci Soc Am J* 65: 527–535.

Stolt, M. H., M. H. Genthner, W. L. Daniels, V. A. Groover, S. Nagle, and K. C. Haering. 2000. Comparison of soil and other environmental conditions in constructed and adjacent palustrine reference wetlands. *Wetlands* 20: 671–683.

Sutton-Grier, A. E., M. Ho, and C. J. Richardson. 2009. Organic amendments improve soil conditions and denitrification in a restored riparian wetland. *Wetlands* 29(1): 343–352.

Titus, J. H. 1990. Microtopography and woody plant regeneration in a hardwood floodplain swamp in Florida. *Bullet Torrey Bot Club* 117: 429–437.

Tweedy, K. L. and R. O. Evans. 2001. Hydrologic characterization of two prior converted wetland restoration sites in Eastern North Carolina. *Transact ASAE* 44: 1135–1142.

Unghire, J. M., A. E. Sutton-Grier, N. E. Flanagan, and C. J. Richardson. 2011. Spatial impacts and wetland restoration on riparian soil properties in the North Carolina Piedmont. *Restor Ecol* 19: 738–746.

USACE and USEPA (U.S. Army Corps of Engineers and U.S. Environmental Protection Agency). 2008. Compensatory mitigation for losses of aquatic resources. 73 Federal Register 19594. Washington, D.C., USA.

van der Valk, A. G., T. L. Bremholm, and E. Gordon. 1999. The restoration of sedge meadows: Seed viability, seed germination requirements, and seedling growth of *Carex* species. *Wetlands* 19: 756–764.

Vivian-Smith, G. 1997. Microtopographic heterogeneity and floristic diversity in experimental wetland communities. *J Ecol* 85: 71–82.

Whisenant, S. G., T. L. Thurow, and S. J. Maranz. 1995. Initiating autogenic restoration on shallow semiarid sites. *Restor Ecol* 3: 61–67.

Whittecar, G. R. and W. L. Daniels. 1999. Use of hydrogeomorphic concepts to design created wetlands in southeastern Virginia. *Geomorphology* 31: 355–371.

Williams, C. B. 1964. *Patterns in the Balance of Nature and Related Problems in Quantitative Ecology*. Academic Press, New York.

Zampella, R. A. and K. J. Laidig. 2003. Functional equivalency of natural and excavated coastal plain ponds. *Wetlands* 23: 860–876.

Zedler, J. B. and R. Langis. 1991. Comparisons of constructed and natural salt marshes of San Diego Bay. *Restoration Manag Notes* 9: 21–25.

Zedler, J. B. and S. Kercher. 2005. Wetland resources: Status, trends, ecosystem services, and restorability. *Annu Rev Environ Resour* 30: 39–74.

Zedler, J. B. and P. H. Zedler. 1969. Association of species and their relationship to microtopography within old fields. *Ecology* 50: 432–442.

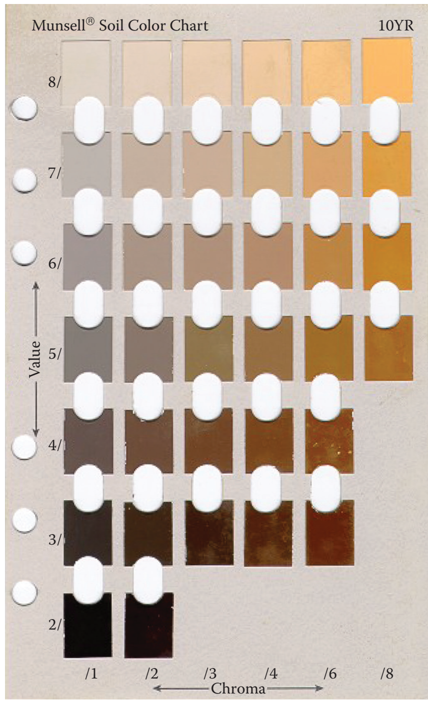


FIGURE 1.3
The 10YR page from the *Munsell® Soil Color Chart*.

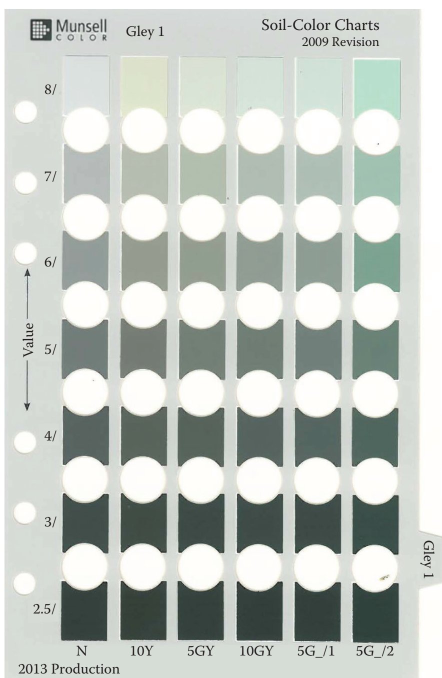


FIGURE 1.4
Example of the gley page from the *Munsell® Soil Color Chart*.

FIGURE 7.2
Example of iron masses (reddish orange colors) between 30 and 50 cm. (Photo provided by John Kelley.)

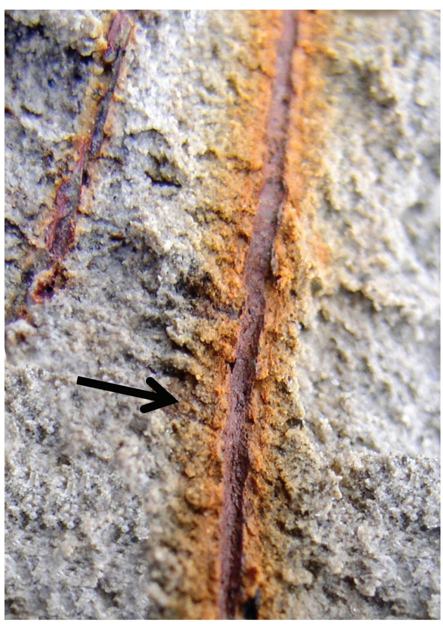


FIGURE 7.3
Example of an iron pore lining (10 mm wide) along a root channel. (Photo provided by John Kelley.)

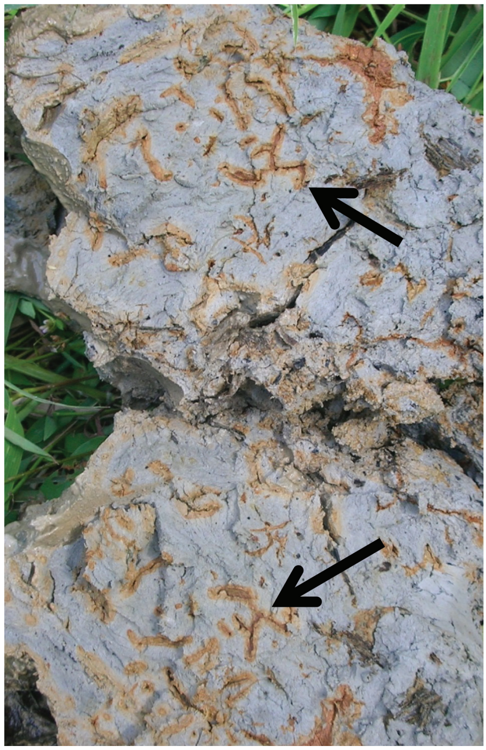


FIGURE 7.4
Examples of iron pore linings (5-10 mm wide) in a horizon containing a Depleted Matrix hydric soil field indicator. (Photo provided by M. Vepraskas.)

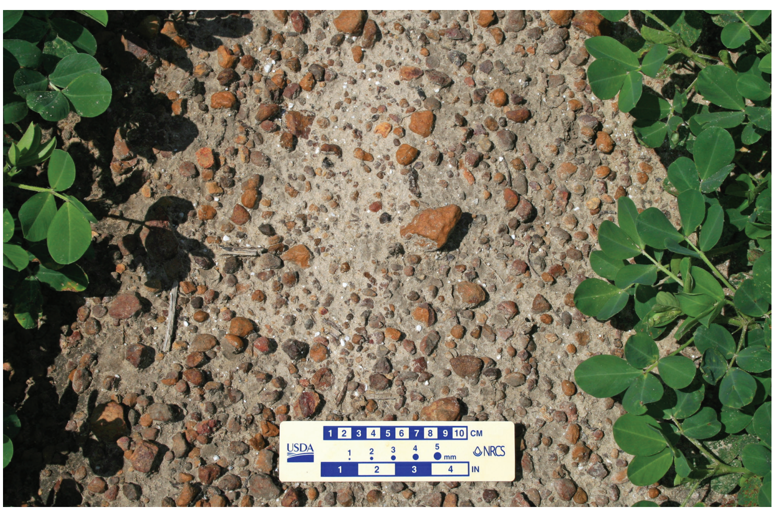


FIGURE 7.5
Iron nodules on the soil surface that were exposed by erosion. (Photo provided by John Kelley.)

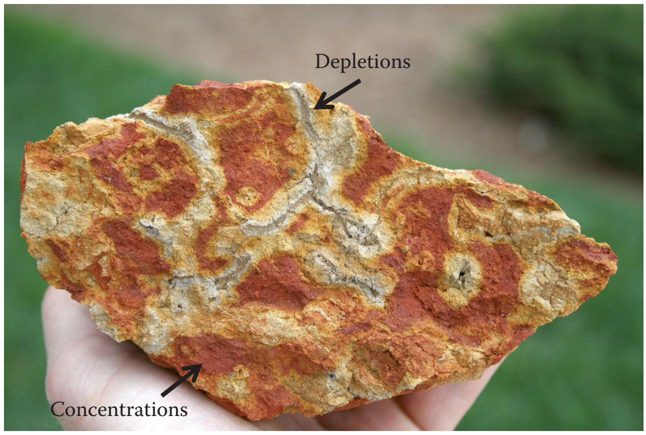


FIGURE 7.6
Gray iron depletions (10 mm wide) along root channels. Reddish orange iron masses in the matrix. (Photo provided by John Kelley.)

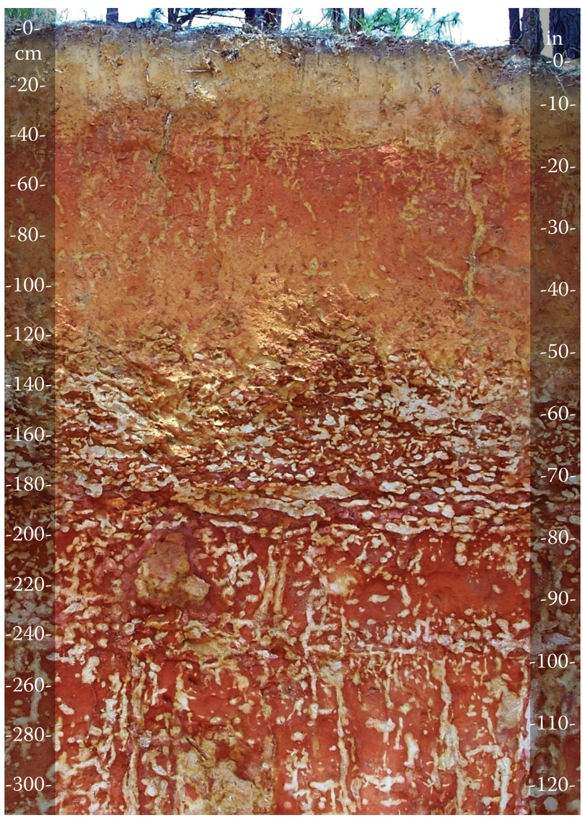


FIGURE 7.7
Iron depletions along root channels primarily below a depth of 120 cm in a Fragic Kandiudult soil in NC. (Photo provided by John Kelley.)

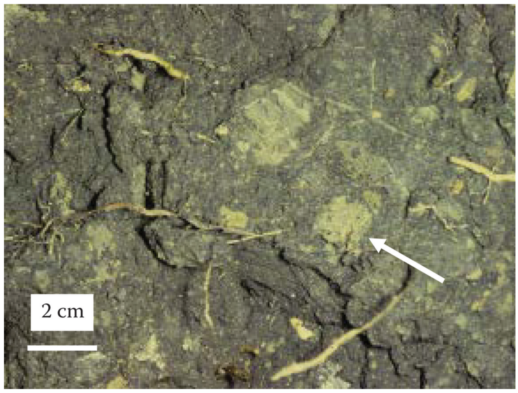


FIGURE 7.8
Gray iron depletions (arrow) in matrix of a silt loam A horizon. (Photo provided by M. Vepraskas.)

FIGURE 7.11
Gray iron depletions (arrows) in a loamy sand E horizon. Example of a Stripped Matrix hydric soil field indicator. Note dime for scale. (Photo provided by Wade Hurt.)

FIGURE 12.10
Example of oxidized iron entering a stream through a seep at valley edge. (Photo from Chowan Co. NC, provided by David Lindbo.)

FIGURE 15.2
Examples of playas near Las Vegas, New Mexico; one dry (left) and one filled with water (right). The playa now filled with water might experience a period of unvegetated soil following the dryout period as much of the perennial grass and other species will have undoubtedly been drowned. The playa that remains mostly unponded will likely remain vegetated and will provide a productive pasture via good run-on and storage of water. (Photo provided by A.J. Miller, NRCS, NM.)

FIGURE 15.4
Salt crust formed by evaporation of water from a floodplain in Salt Creek of the Tularosa Basin in southern New Mexico. The gypsum salt crust is commonly coated on the underside by halophytic algae. (Photo by G. Cates, NRCS, NM.)

FIGURE 15.5
A Typic Aquisalid in the Carrizo Plain, California. In this playa landform position, small amounts of organic matter can accumulate in the playa surface, enough to feed anaerobic soil microbes. Notice the low chromas of the surface horizons (between ribboned pins) indicative of redoximorphic conditions. Also visible in the first horizon is finely disseminated secondary salts that may mask the colors needed to identify hydric indicators. This soil does not have enough organic matter at the surface to form a distinct A horizon; directly below the salt crust is a Cz1 horizon. (Photo by A. J. Miller NRCS, NM.)

FIGURE 15.6
Gypsic Aquisalid observed in the Tularosa Basin of New Mexico. In a layer just below the surface, darkly colored Mn^{2+} accumulations, likely as manganese monosulfide (MnS), form as mineral coatings around ped surfaces, an indicator that both dissolved manganese and a source of sulfur exist. Precipitates of this reduced form of Mn can be easily verified in the field by applying 3% H_2O_2 and watching the minerals dissolve rapidly, leaving no trace of the black MnS. (Photo by A. Miller, NRCS, NM.)

FIGURE 15.7
This ped sample from the Stanley series shows a combination of two materials due to vertic mixing. The light greenish color (5YR 8/1) is from reduction in the soil profile, and the redder hues (5YR 7/2) are from material that is washed or blown into the profile when the soil is dry and deeply cracked. (Photo provided by A. Miller, NRCS, NM.)

FIGURE 18.1
Development of an organic-rich surface layer 20 years following marsh creation. (Photo provided by C.B. Craft.)

FIGURE 18.5
Aerial photographs of the same creek network at Pond A21 in April 2008 (2 years post breach), September 2009 (3.5 years post breach), and June 2011 (5 years post breach) illlustrating the rapid sedimentation and natural re-vegeatation of the marsh. (Photo by Chris Benton.)

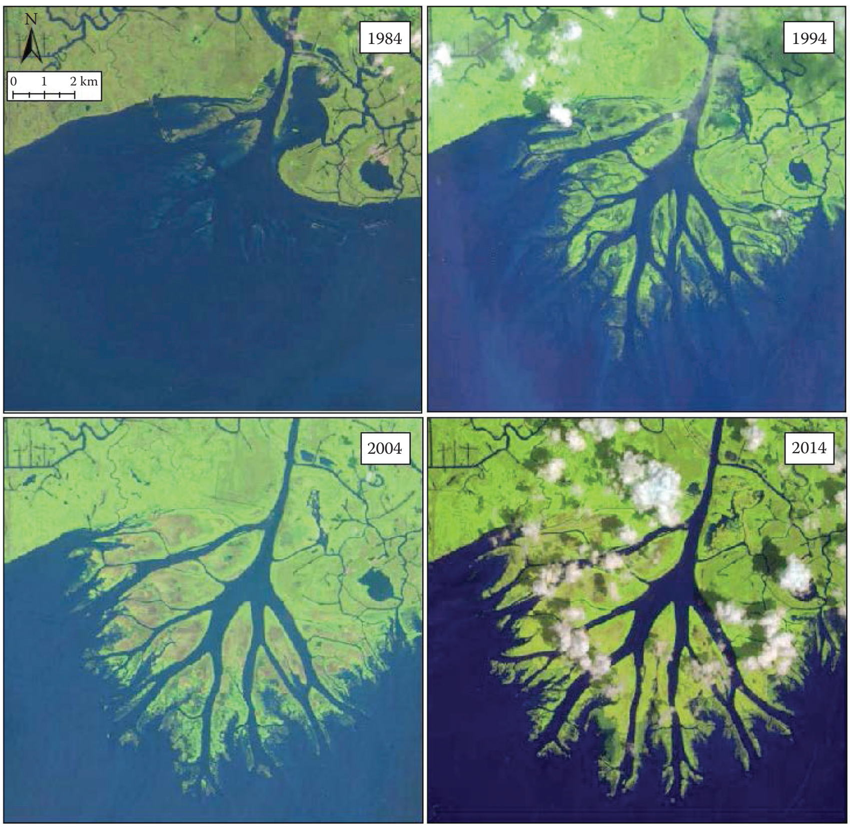


FIGURE 18.6
Development of the Wax Lake Delta, Louisiana, over a 30-year period. The delta formed as the result of an unplanned driver diversion to reduce flooding along the Atchafalaya River in the 1970's. Photos were taken by Landsat 5 (1994–2004) and Landsat 8 (2014) satellites.

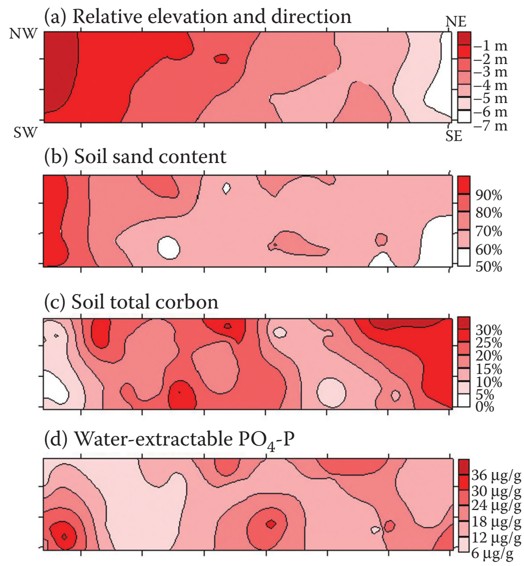


FIGURE 19.3
(a–d) Distribution of selected soil parameters across the restored Carolina bay wetland in Cumberland Co., NC. Elevation measurements are referenced to the northwest (upper left) corner.

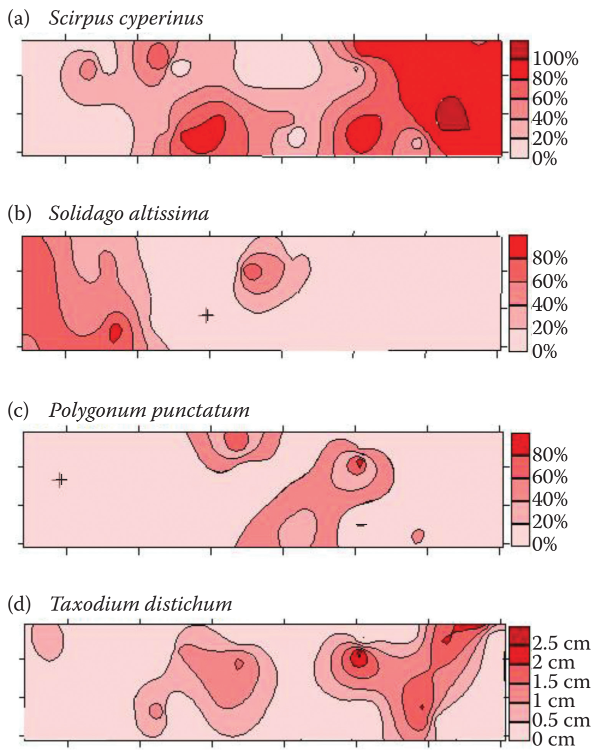


FIGURE 19.4
(a–c) Distribution of the three most commonly observed species at the restored Carolina bay wetland in Cumberland Co., NC. (d) Distribution of planted *Taxodium distichum* saplings.

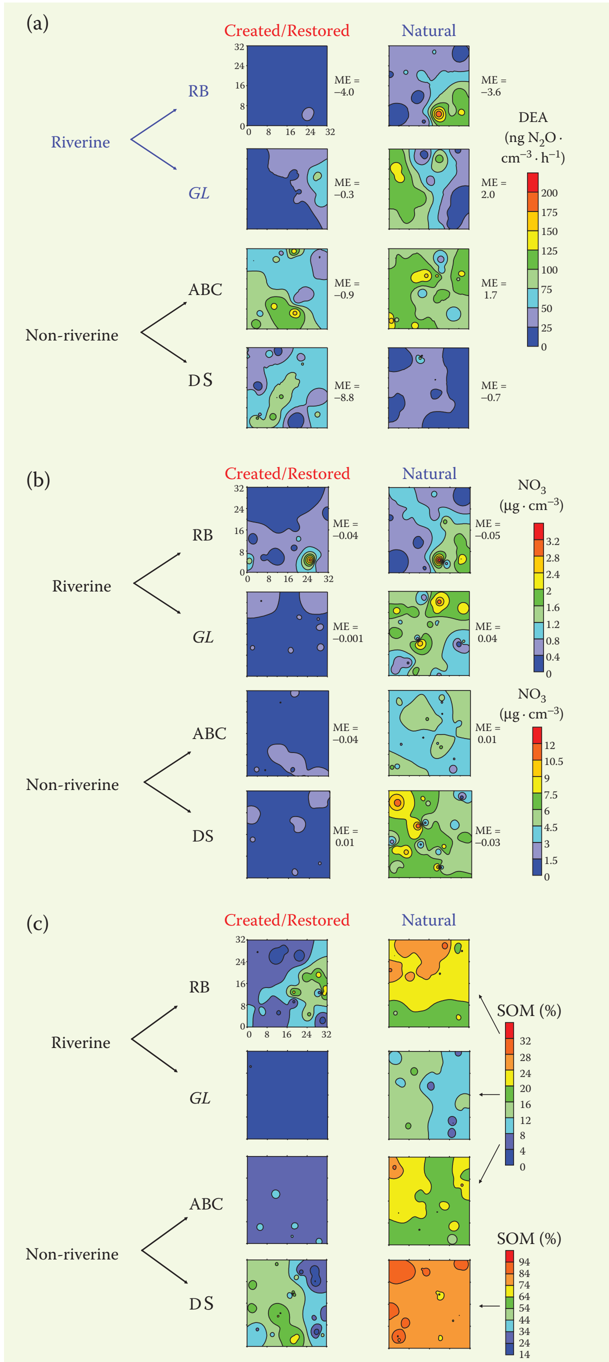


FIGURE 19.5
(a) Spatial distribution of predicted denitrification enzyme activities (DEA), (b) spatial distribution of nitrate (NO_3-N), and (c) spatial distribution of soil organic matter (SOM) in restored and natural wetlands. Sites consist of Rowel Branch (RB), Grimesland (GL), ABC, and Dismal Swamp (DS). (Adapted from Bruland, G. L. and C. J. Richardson. 2006. *Wetl Ecol Manag* 14: 245–251; Bruland, G. L. and C. J. Richardson. 2005a. *Soil Sci Soc Am J* 69: 273–284.)

ENVIRONMENTAL ENGINEERING

Second Edition

WETLAND SOILS

Genesis, Hydrology, Landscapes, and Classification

"This is an exceptional reference … a virtual encyclopedia on hydric soils written by the nation's leading soil scientists offering readers a complete understanding of hydric soils, including the processes responsible for their development, their diagnostic properties, and how to delineate hydric soils in the field."

—Ralph Tiner, Institute for Wetland & Environmental Education & Research, Inc.
University of Massachusetts, Amherst, USA

"The main strength is the comprehensiveness of the book. It covers all aspects of wetland soil characteristics including chemistry, hydrology, morphology, biota, classification, identification and delineation. … written in a way that anybody can understand and therefore it is useful to have it at hand."

—Jan Vymazal, Czech University of Life Sciences, Prague

"… an accessible, soil-centric perspective on the hydrologic, physical, chemical and biological phenomena that give rise to these enigmatic ecosystems. … stays true to the goal of providing students and practitioners with the basic science theory they need to understand wetland soils, and insights on how the science applies to practical applications."

—J. Patrick Megonigal, Smithsonian Environmental Research Center, Edgewater, Maryland, USA

"…offers a comprehensive coverage of wetland soils with great detail provided about their physical, chemical and morphological properties while simultaneously providing a strong national level review of their landscape associations. … maintains the essential sequencing of the original version with basic fundamental and underpinning topics being covered first, followed by applications, and then field landscape distributions … will continue to be the standard reference and textbook for both undergraduate and graduate courses and will also serve as an invaluable background resource for wetland professionals working in the field."

—W. Lee Daniels, Department of Crop & Soil Environmental Sciences, Virginia Tech, Blacksburg, USA

K14282

ISBN: 978-1-4398-9698-3

90000

9 781439 896983

CRC Press
Taylor & Francis Group
an informa business
www.crcpress.com

6000 Broken Sound Parkway, NW
Suite 300, Boca Raton, FL 33487
711 Third Avenue
New York, NY 10017
2 Park Square, Milton Park
Abingdon, Oxon OX14 4RN, UK

www.crcpress.com