

AN ABSTRACT OF THE THESIS OF

David V. D'Amore for the degree of Master of Science in Soil Science
presented on July 5, 1994. Title: The Stratigraphy, Hydrology, and
Redoximorphic Character of the Jackson-Frazier Wetland.

Abstract approved:  **Redacted for Privacy**

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Transitional areas between upland and aquatic habitats, commonly known as wetland, were once viewed as unproductive areas and were drained for farming or pasture. Wetlands are now accepted as significant ecological resources, and their protection is a mandate of federal, state, and local land managers. Due to the diversity of wetland areas, the appropriate assessment of wetland resources cannot be accomplished without long term monitoring of wetland functions. Knowledge of the duration of saturation and associated anaerobic conditions of soils in wetlands is critical to correctly classify and assess wetland areas.

Soil, hydrological, and biogeochemical characteristics of the soils of the Jackson-Frazier wetland were observed from October 1992 through March 1994. Weekly observations of water levels and redox potential at depths of 25, 50, and 100 cm were made in order to characterize the degree and duration of saturation and the anaerobic conditions in the soil over time.

Permanently installed piezometers measured free water in the soil and indicated the presence of two separated water tables from the onset of the rainy season in October until February when the entire soil profile became saturated with free water. Platinum electrodes measured redox potential in the soil and indicated

anaerobic conditions for ten months during the first season of observation and through March of the second season. Anaerobic conditions were considered to be achieved when Fe^{+3} was reduced to Fe^{+2} at a potential of 200 millivolts. The highly reducing conditions correspond to periods of soil saturation indicated by piezometers. Concentrations of iron and manganese observed in soil profiles correspond to conditions of prolonged saturation and reduction confirmed by monitoring.

A soil stratigraphic study done with auger holes revealed a recent alluvial deposit of montmorillonitic clay overlying lacustrine silts identified as the Irish Bend Member of the Willamette Formation. The clay deposit overlying the surface of the wetland acts as an aquitard and creates extensive surface ponding, which maintains the saturated habitat required for wetland vegetation. The subsurface hydrology is controlled by water flowing through the Irish Bend silts which results in saturation of the soils from below. Biogeochemical transformations of iron and manganese due to suboxic and anaerobic conditions are controlled by this type of soil saturation in the Jackson-Frazier wetland.

**The Stratigraphy, Hydrology, and Redoximorphic Character
of the Jackson-Frazier Wetland**

by

David V. D'Amore

A THESIS

submitted to

Oregon State University

**in partial fulfillment of
the requirements for the
degree of**

Master of Science

Completed July 5, 1994

Commencement June 1995

APPROVED:

Redacted for Privacy

Professor of Soil Science in charge of major

Redacted for Privacy

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Date thesis is presented July 5, 1994

Typed by David D'Amore for David D'Amore

ACKNOWLEDGMENTS

The time I have spent at Oregon State University during my graduate education has been very rewarding and productive due to many friends, colleagues, and advisors who shared classes, seminars, field work and informal gatherings with me over the past two years.

My thanks and appreciation first go to Will Austin who introduced me to wet soil monitoring, taught me the intricate details of platinum electrode construction and endured many long discussions about redox theory. I also thank fellow students and friends, Scott Stewart and Lisa Seter, for their help with field work and for their interesting discussions of the chemistry and mineralogy of wet soils. I thank Peggy Erickson for her hard work on compiling data and coaxing me over to windows.

I thank my committee members, Dr. John Baham, Dr. Robert Frenkel, and Dr. Jim Boyle for reviewing my thesis and serving on my committee. I wish to extend my deepest appreciation and thanks to my advisor, Dr. Herb Huddleston, who gave me the opportunity to undertake this project and whose guidance and careful review of my work made the final product possible. I also thank the USDA Soil Conservation Service for their support of the wet soils project and express my hope that this research continues in the future.

Finally, a special thanks to Toni DeSanto who offered many hours of her time sharing her experience on thesis organization and editing, as well as her emotional support throughout the process of obtaining my degree.

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The Stratigraphy, Hydrology, and Redoximorphic Character of the Jackson-Frazier Wetland

Chapter I. Introduction

Transitional areas between upland and aquatic habitats, commonly known as wetland (Mitsch and Gosselink, 1989), were once viewed as unproductive areas and were drained for farming. Wetlands are now accepted as significant ecological resources, and their protection is a mandate of federal, state, and local land regulations. Due to the diversity of wetland areas the appropriate assessment of wetland resources cannot be accomplished without monitoring of wetland functions. Knowledge of the duration of saturation and associated anaerobic conditions of soils in wetlands is critical to correctly classify wetlands and assess their function.

In 1991, Oregon State University, in cooperation with the USDA Soil Conservation Service, started a wet soil monitoring program in Oregon. The goal of this project was to investigate the different patterns of soil saturation that occur in several soils with different drainage classes, climates, and parent materials. Originally, the project focused on two catenas in the Willamette Valley (Austin, 1994). The project was expanded to include arid, volcanic derived soils in Bend, and udic, iso-mesic soils in coastal Tillamook County. The Willamette Valley sites were also expanded to include three plots on what were believed to be some of the wettest soils on the landscape in the Jackson-Frazier wetland.

The Jackson-Frazier wetland is recognized as a significant and valuable natural resource. It is home to several important remnant Willamette Valley prairie plant communities. The wetland also serves as viable habitat for several species of birds, beavers, and deer as well as small mammals. In 1992, the wetland became a Benton County park and is managed primarily for the protection of the wetland

resources and serves as a multi-purpose area. One of the goals of the management plan is the coordinated research of wetland characteristics and functions.

The results of an investigation of the soil stratigraphic, hydrologic, and biogeochemical characteristics of the soils of the Jackson-Frazier wetland as observed from October 1992 through March 1994 are reported in this thesis. The soil characterization study was intended to expand the OSU-SCS wet soil monitoring project as well as complement a previous study of the distribution and vegetation ecology of the Jackson-Frazier wetland (Marshall, 1985). The goals of the present project were:

- 1) To map and define the soils of the wetland and their associated stratigraphy and assess the influence of these factors on the wetland functions.
- 2) To observe and measure the subsurface hydrology and characterize the type of saturation dominant in the soils.
- 3) To measure the oxidation and reduction (redox potential) of the soils to determine the extent and intensity of anaerobic conditions.
- 4) To observe and record the occurrence of redoximorphic features in soils and relate them to measured hydrological and redox potential conditions.

Chapter II. Geomorphology, Stratigraphy, and Soils of the Jackson-Frazier Wetland

Introduction

Soil/geomorphic associations provide a framework for studying soils with specific characteristics to evaluate landscapes. Pedology addresses the processes of soil formation, while geomorphology is concerned with the arrangement and differentiation of landforms and the processes that have been or are shaping them (Gerrard, 1992). Linking these two areas of study provides a cohesive approach to identifying and understanding soils in the landscape. A 'geomorphic surface' has been defined by Daniels and Hammer (1992) as "*an area of land that has a development history related in space and time*". Soils identified on specific geomorphic surfaces have the potential to accurately record the history of landscape development, as soils are intimately linked to the dynamic equilibrium of the landscape (Gerrard, 1992). Daniels and Hammer (1992) call soil "integrators" of their past and present environment.

In order to interpret geomorphic surfaces and soils associated with them, one must establish some basic relationships based on geologic history, past and present climate, vegetation, source material mineralogy, and weathering patterns of an area. Landscape features are often closely related to factors that influence soil development and can be used to predict soil types. The relationship between soils and geomorphic surfaces can be used to make inferences about the structure and function of an area, interpret the possible modes of formation of soils, and establish sites for detailed soil investigation.

The goals of the stratigraphic part of this research project were to identify the soil stratigraphic units that comprise the wetland, determine their extent, and determine their influence on the hydrology of the wetland.

Geomorphology of the Willamette Valley

Pleistocene flood events have created a complex stratification in the soils of the Willamette Valley of Oregon (Baldwin, 1964). Relationships between stratigraphic deposits and soils of the Willamette Valley were first elucidated by Balster and Parsons (1969). They evaluated the geomorphology of the valley and correlated soils with geomorphic surfaces.

The Willamette Valley is filled with recent (i.e. holocene) alluvium consisting of fine textured material overlying fine-silty and silty sediments of pleistocene age. The pleistocene material was first identified and named the Willamette silt by Allison (1953). The silty deposit was later renamed the Willamette Formation and divided into four distinct stratigraphic units by Balster and Parsons (1969): the Wyatt, Irish Bend, Malpass, and Greenback members. The Irish Bend member is described as a thick deposit that consists of "faintly bedded, micaceous, silty sediments with well defined upper and lower boundaries" (Balster and Parsons, 1969). Balster and Parsons hypothesized that these silts may have been locally derived, but more recent research has shown the silts to be of glacial origin (Glasmann and Kling, 1980).

Sedimentation on the valley floor can be very diverse. Terraces under which there is a clay stratum exhibit surface ponding at shallow depth beneath coarser sediments. This clay deposit is not a feature of the entire region, but occurs on terraces which have not been subject to erosion or sedimentation. The clay deposit

is the feature known as the Malpass clay. The Malpass member is a very fine deposit derived from lacustrine sediments of the mid-pleistocene. Overlying the malpass is the Greenback member, which consists of silty lacustrine material of the late Pleistocene.

Problem Statement

Soil morphological interpretations are a particular challenge in areas of varied geologic history such as the Jackson-Frazier wetland. The physical and chemical processes that occur in a soil are expressed as morphological features which are in a soil profile and are used in soil classification to differentiate soils in the field. Soils formed from alluvial parent materials often display stratification within the soil profile of a pedon. However, designation of horizons that have developed in different source materials as lithologic discontinuities is contingent on distinguishing between source materials in which each soil horizon formed. Positive field identification of lithologic discontinuities is often difficult, and the field soil scientist is limited to tentative conclusions about the presence of different parent materials based on field observations of texture, structure and mineralogy. However, once an interpretation of a soil's origin is made it can be extrapolated to the larger landscape.

Although specific geomorphic associations were established by Balster and Parsons (1968) for the Willamette Valley and associated foothills, these relationships must be re-evaluated to confirm or expand existing soil-geomorphic associations. Some revisions to definitions of the original surfaces have already been made. The Bethel surface was identified by Gelderman (1970), who subdivided low rounded hills from 3 to 12 meters in relief into a distinct geomorphic unit. Glasmann et al. (1980)

defined the Brateng surface, which established a new soil geomorphic association from 80 to 122 meters elevation on ridge spurs.

The drainageways of the foothills surrounding the valley have cut through older surfaces and formed younger depositional surfaces on which there is a new mosaic of poorly drained soils in depressions. Poorly drained channel areas are surrounded by better drained terraces and footslopes on the western margin of Benton County. The Jackson-Frazier wetland is located on the poorly drained soils of the Ingram surface (Balster and Parsons, 1968) on an alluvial fan of the Jackson and Frazier creek drainages. The poorly drained soils of the channels and fans of Coast Range streams are important sources for wetlands. The deposition of sediments on these surfaces leads to the creation or expansion of wetlands on the western margins of the Willamette Valley. Correctly classifying the stratigraphy of the Jackson-Frazier system could lead to identification of similar areas that might be critical wetland formation areas.

A thorough understanding of the stratigraphy of the source materials for the wetland can also help explain the control the sediments play in the hydrology and biogeochemistry of the wetland. Inferring the processes responsible for the formation of the soils of the wetland is an important part of a complete pedological evaluation, and making inferences about processes requires a complete evaluation of the hydrology, stratigraphy and geochemistry of the soils.

Methods

Site Description

The Willamette Valley is a broad, level area lying between the Cascade Range and the Coast Range in Western Oregon. The Jackson-Frazier wetland is a 64 hectare area located at the base of the foothills of the Coast Range on the northern edge of the town of Corvallis. The wetland has forested, shrub-scrub, and prairie palustrine plant communities (Marshall, 1985). Low relief and the presence of numerous beaver dams below the confluence of the Jackson and Frazier creeks has created an area of complex overland flow and semi-permanent saturation. The wetland is drained to the south by Stewart Slough and to the north by Frazier Creek ditch. These drainages are tributary to the Willamette River two miles to the east. The wetland lies at approximately 66 meters in elevation.

Soil Morphology

Five transects were established from the southernmost boundary of the wetland to the northern boundary using compass bearings. Thick vegetation made it necessary to lay out transects in a zig-zag pattern, and to space points at unequal distances, but as close to 50 meters as possible. The location of each point on the transect was determined by measuring the compass bearing and measuring the distance from the previous point. Elevations at each point were determined with a transit and stadia rod.

A square grid was drawn over the area of the wetland for preparation of a topographic configuration. Elevations at grid intersections were taken from the closest adjacent sample point or the average of points if between points. Some

points on the exterior border were extrapolated from field reconnaissance. Plots for relative elevation were completed by establishing a datum from city of Corvallis survey information which was used to reference field data. All values are reported in meters.

Soil core samples were taken at 50 meter intervals along the transects with a bucket auger. Observations of color changes, (a Munsell color chart was used to describe color patterns), mottling patterns and textural changes were made and depths to each feature recorded. Observations were made to a depth of two meters.

Soil profile descriptions were completed on three pits using standard soil morphology nomenclature. (Soil Survey Division Staff, 1993).

Mineralogy

Four sites (transect point numbers 48, 85, 129, and 109) were chosen along an elevational gradient from 0 to 2 meters within the wetland for detailed mineralogical analyses. At each site, soil samples were taken with a soil auger from the A, B, and C horizons at depths of 50, 100 and 180 cm, respectively. Samples were placed in sealed bags to retain field soil moisture content until the analysis was done.

A subsample of each bulk sample was removed and mixed with 150 ml distilled water and 5 ml dispersant (NaHMP-sodium hexametaphosphate) and shaken for eight hours. After dispersion, the sand and silt ($> 2 \mu\text{m}$) and clay ($< 2 \mu\text{m}$) fractions were separated by centrifugation. The sand and silt fraction was discarded. The clay fraction was saturated with magnesium by washing three times with 0.5M MgCl_2 followed by three rinses with distilled water. Slides were then prepared for X-ray diffraction (XRD) analysis by the paste method (Thiessen and Harward, 1962).

The remaining clay fraction was then saturated with potassium by washing three times with 1M KCl. Clay minerals were analyzed using a Norelco Model 3100 CuK alpha monochromatic instrument with a compensating slit. Slides were analyzed at one second intervals each 0.2 degrees 2-theta spacing at 40 Kv and 35 ma power. Following the initial analysis, the magnesium saturated slides were then solvated with ethylene glycol and X-ray diffraction patterns were obtained using the method described above.

Results

Soil Morphology

Profile Descriptions

Soils described at pit sites I and II were classified as Bashaw soils. The Bashaw series is characterized by a thick, clayey mollic epipedon that has many large vertical cracks extending to the soil surface. The slow vertical drainage and shrink-swell characteristics of the soil result in a coarse prismatic structure. The Bashaw soils at the Jackson-Frazier sites are underlain by silty sediments which have few fine to common fine prominent strong brown (7.5YR 4/4) iron masses.

The soil described at pit site III has been tentatively identified as the Cove series. This soil has a thick mollic epipedon, and clay textures extend to a depth of almost two meters. Vertical cracks between prisms are apparent but do not extend to the surface in this soil. There are many medium and many fine black (N 2/0) iron-manganese concretions in the Bss horizons, but no brown iron masses. Silty sediments like those at plots I and II were not observed within the depth of sampling.

Topographic Survey

The topographic surface of the wetland is displayed in figure 1. The wetland slopes upward from a low point of 65 meters in elevation in the southeast to a high point of 68 meters in elevation in the northwest. From southwest to southeast the wetland is essentially flat with elevations between 65 and 66 meters. This surface has the configuration of an alluvial fan whose source streams enter the wetland in the northwest corner and spread out over the surface down the elevation gradient to outlet channels located in both the southeast corner and eastern edge of the wetland.

Soil Auger Sampling

Soil cores taken from throughout the wetland reveal three major stratigraphic units: a black (N 2/0) clay over an olive brown (2.5Y 4/3) silty clay, over a light olive brown (2.5Y 5/3) silty clay loam. Soils in the southeast quadrant of the wetland have the thinnest layers of black clay, and the contact with the light olive brown silty clay loam deposit is at a depth of 100 to 150 cm (figs. 2, 3, 4). Prominent red mottles appear 10 to 20 cm above the light olive brown (2.5Y 5/3) silty clay loam material. The appearance of the light olive brown silty clay loam material and the presence of red mottles above the deposit corresponds to the Irish Bend silt stratigraphic deposit described by Balster and Parsons (1968).

The wetland elevation drops to a low point in the southeast corner in the middle of a vernal pond that may at one time have been excavated (Marshall, 1985). Here, the contact with the Irish Bend silt is shallowest, 100 to 130 cm. Depth to the Irish Bend silt contact increases to 160 cm toward the northeast boundary.

Figure 1. Topographic surface of the Jackson-Frazier Wetland

Figure 1.

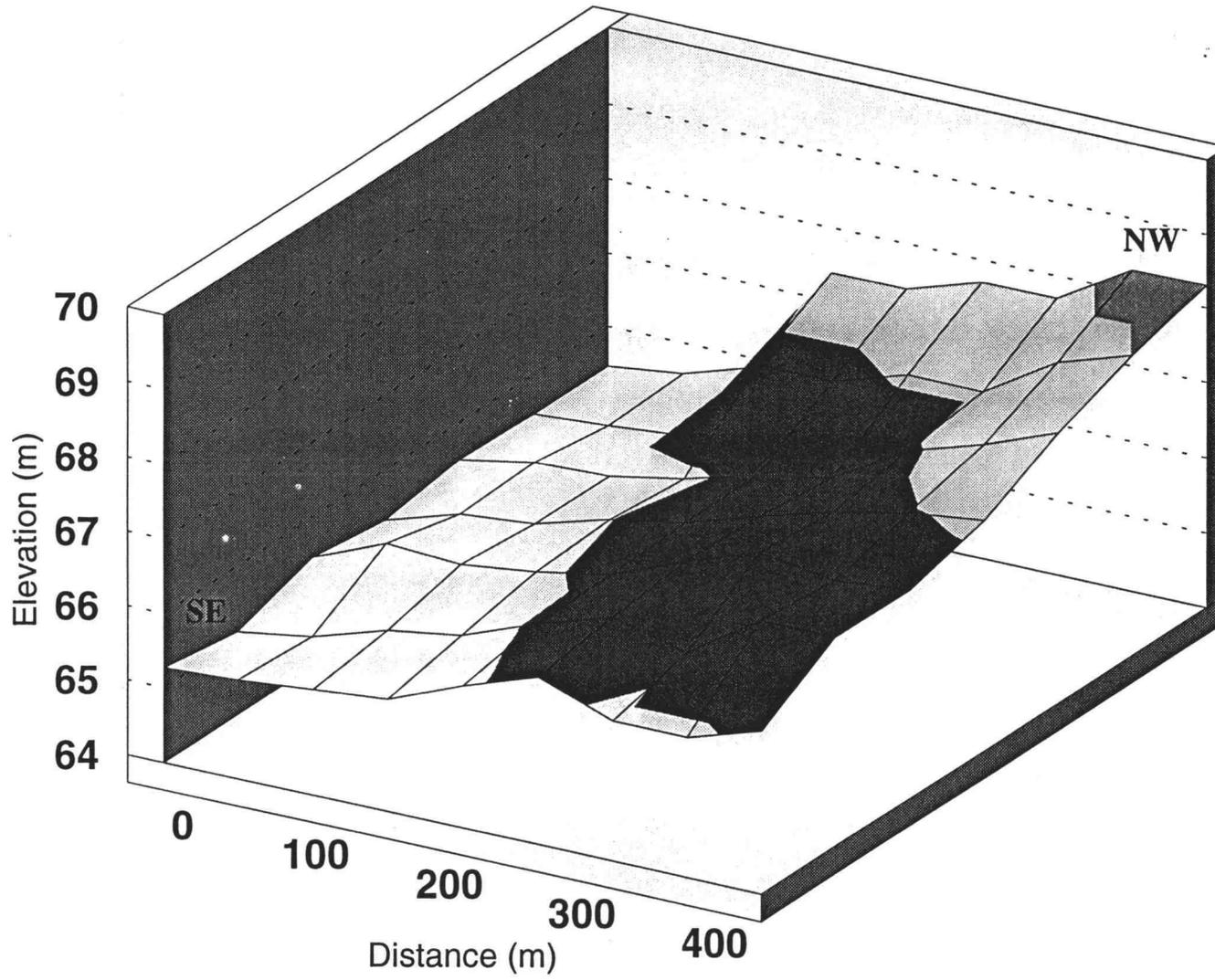


Figure 2. SE to NW stratigraphic transect in Jackson-Frazier Wetland

Figure 2.

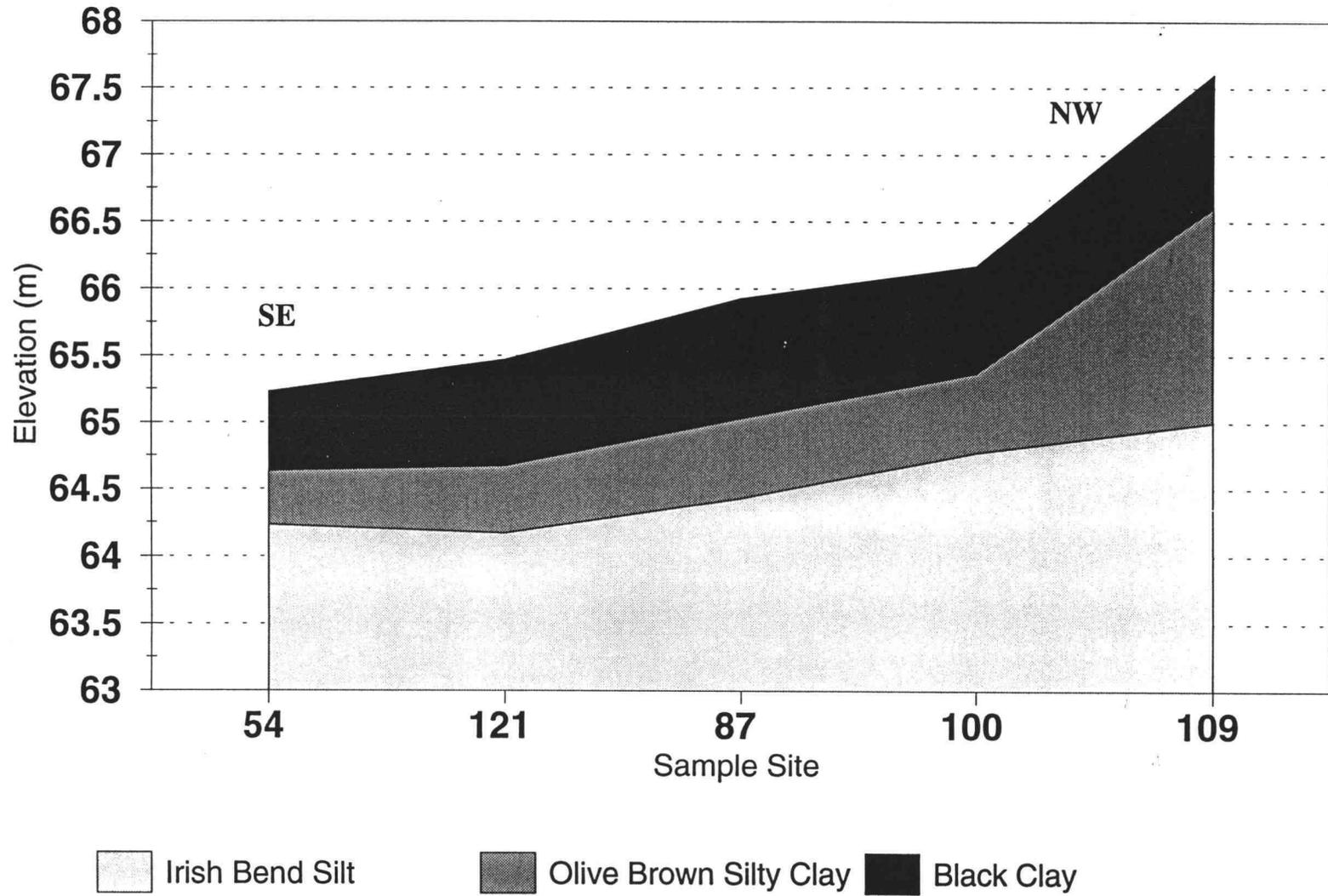


Figure 3. SW to NE stratigraphic transect in Jackson-Frazier Wetland

Figure 3.

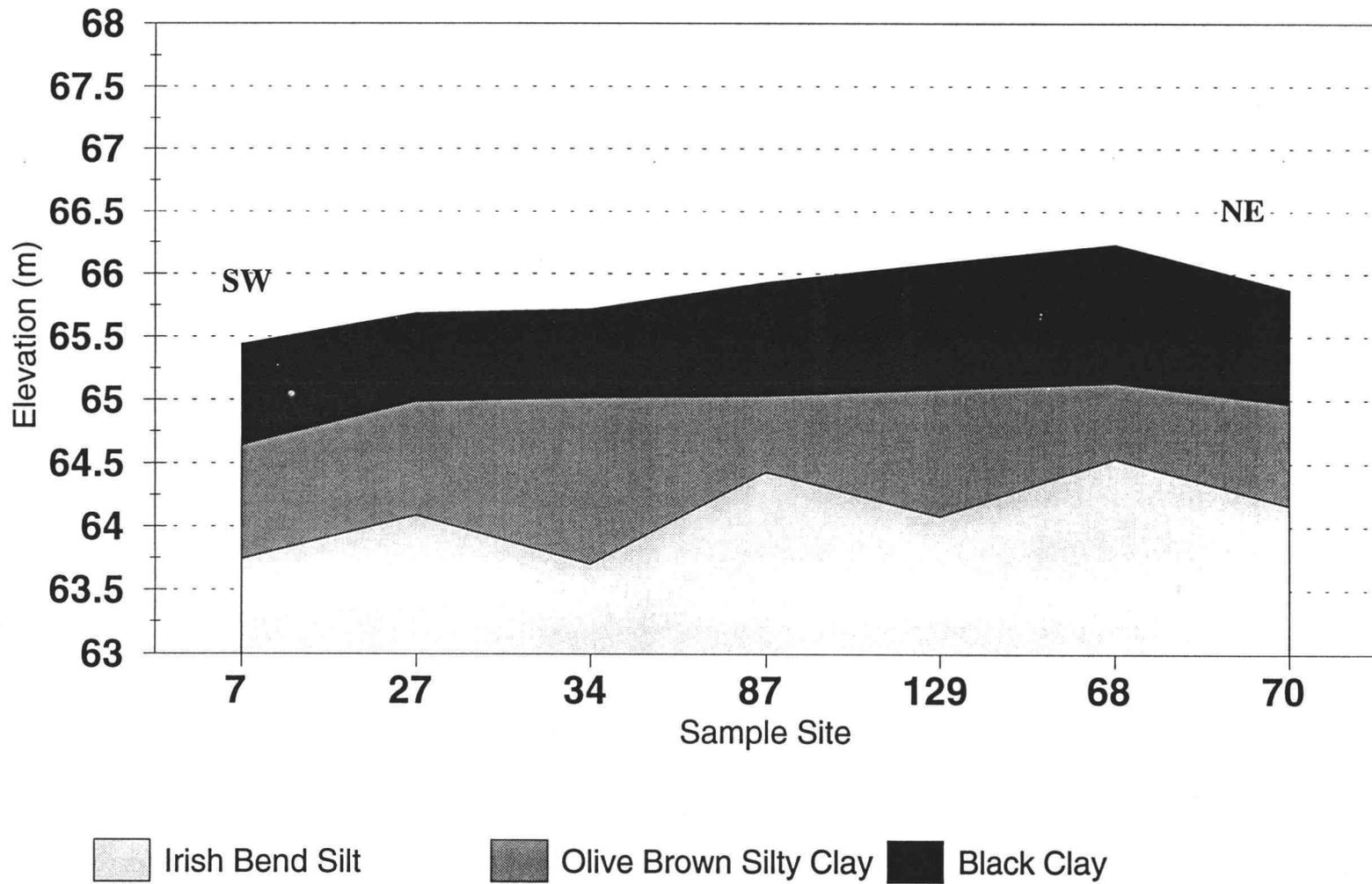
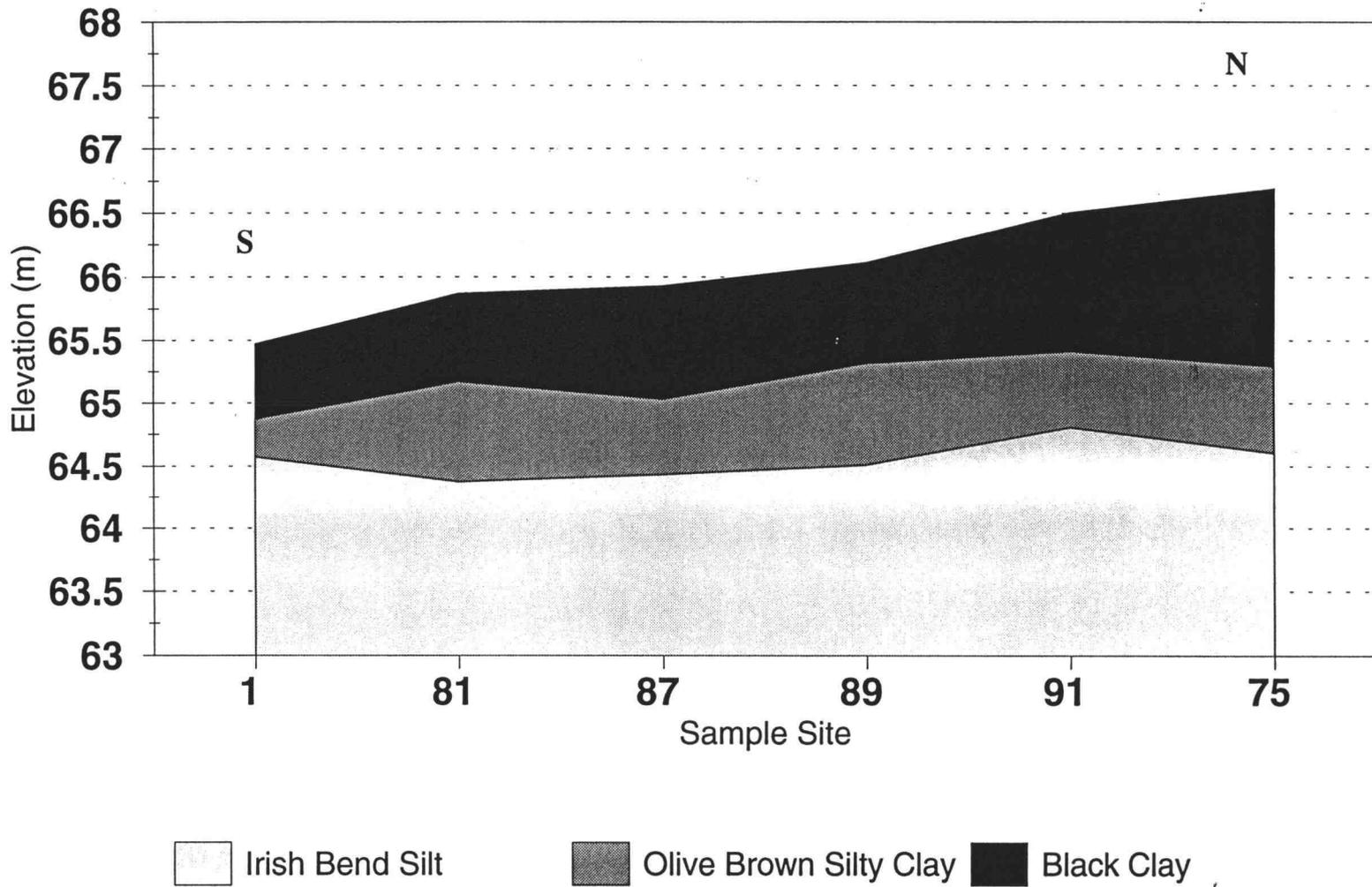


Figure 4. S to N stratigraphic transect in Jackson-Frazier Wetland

Figure 4.



Moving northwestward from the vernal pond, the black clay horizon thickens to a depth of 180 cm over the olive brown silty clay layer. The northwest quadrant has the thickest clay deposit and corresponds to the Cove soil profile description. In this area the contact with the underlying olive brown silty clay and Irish Bend silt is greater than 2 meters deep.

The clay horizon remains greater than 180 cm thick from the northwest corner along the western margin of the wetland, tapering only slightly to silty clay and Irish Bend silt material at 170 cm into the southwest quadrant. The Irish Bend silt deposit is encountered once again in the southeast quadrant where the black clay and silt clay deposits thin yielding a contact with the Irish Bend silt at 130 cm.

Clay Mineralogy

Locations of the four sites sampled for mineralogical analysis are shown in figure 5. The X-ray diffraction patterns (XRD) were nearly alike from the samples taken at the 50 and 100 cm depths (figs. 6, 7) and at the 180 cm depths at sites 109 and 129 (fig. 8). The patterns showed smectite, vermiculite and kaolinite. The absence of clay size mica is apparent from the 10 Å area on the Mg-glycol pattern. This is the area above the 9°, 2-theta mark on the X-ray diffraction pattern.

The 10 Å peaks (9°, 2-theta) on the Mg-glycol patterns from the 180 cm depth at sites 48 and 85 (fig. 8) clearly show a different mineral assemblage than the samples from the 50 and 100 cm depths at these sites (figs. 6 and 7). These patterns include more mica than the samples from 50 and 100 cm, which is evident at 10 Å (9°, 2-theta) on the Mg-Glycol pattern.

Figure 5. SE to NW mineralogy transect in Jackson-Frazier Wetland

Figure 5.

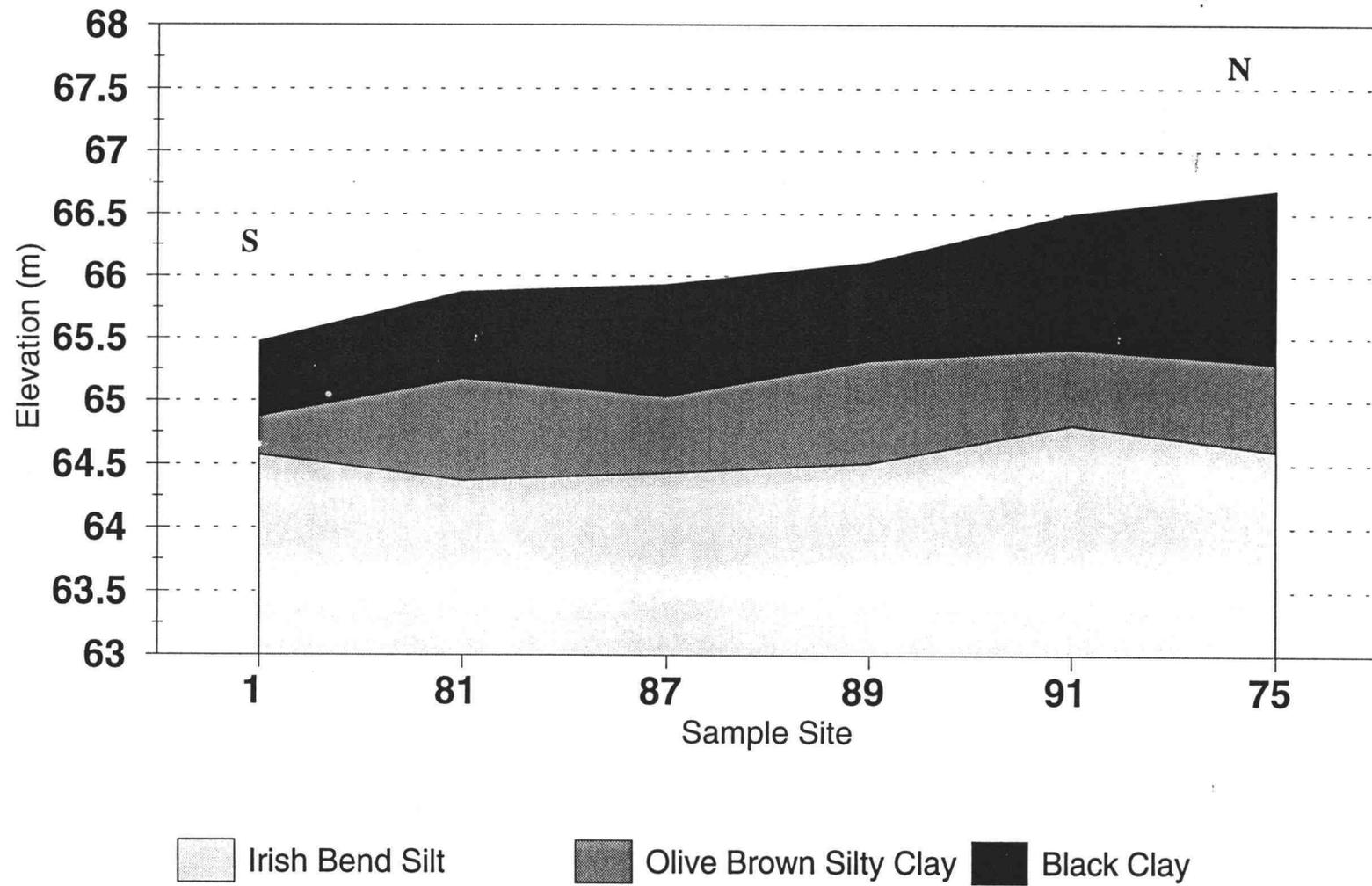


Figure 6. XRD patterns of soils in SE to NW transect at 50 cm

Figure 6.

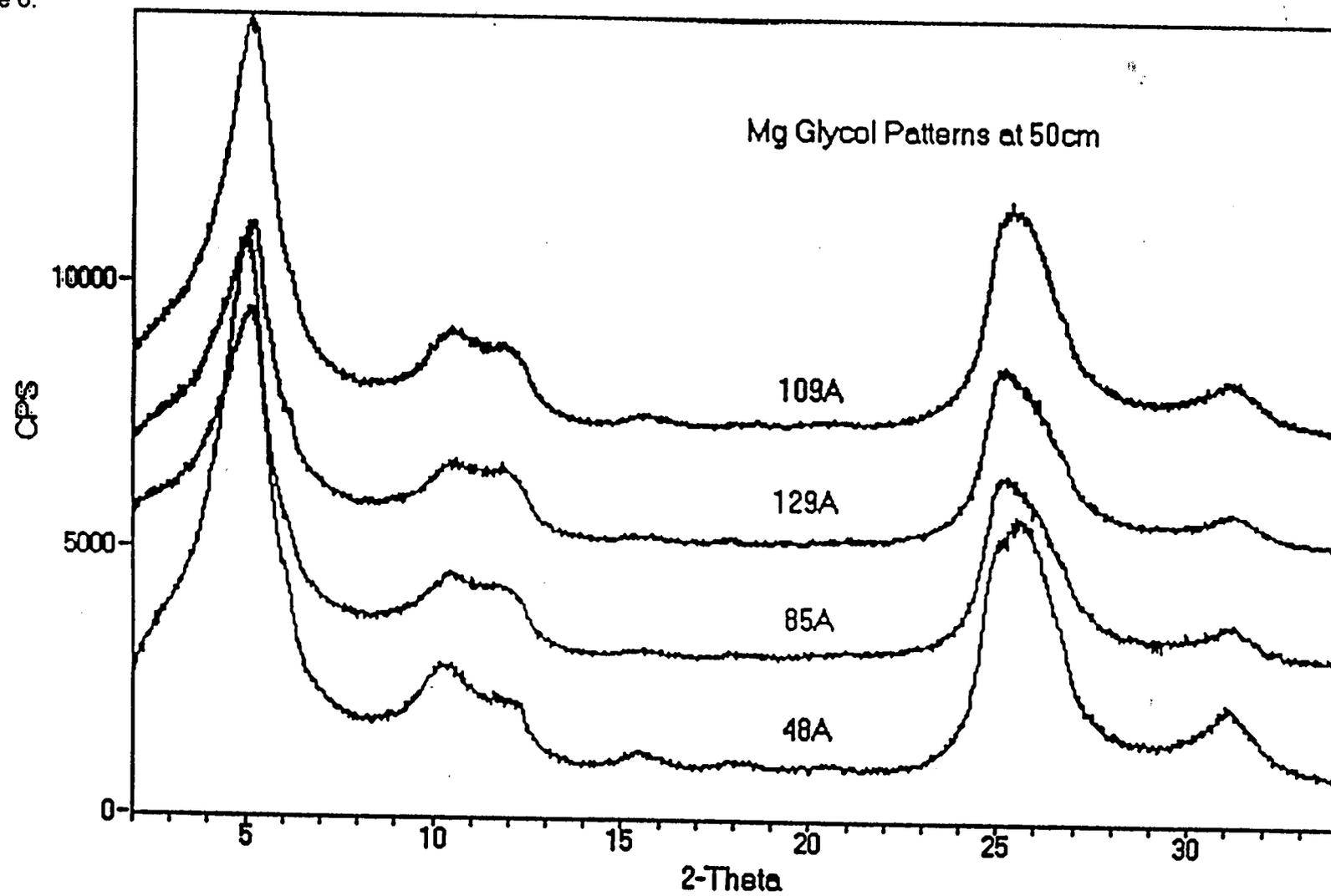


Figure 7. XRD patterns of soils in SE to NW transect at 100 cm

Figure 7.

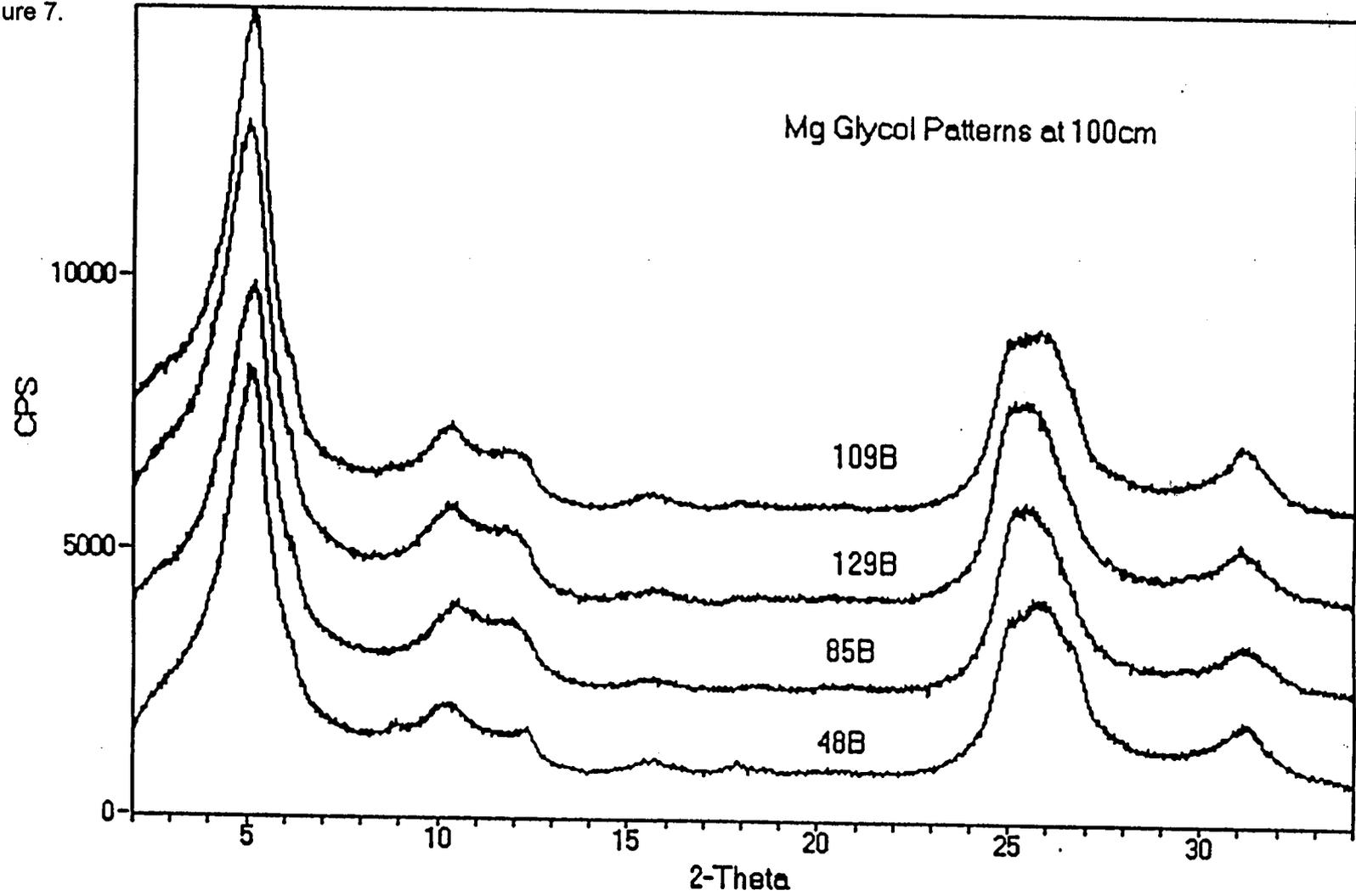
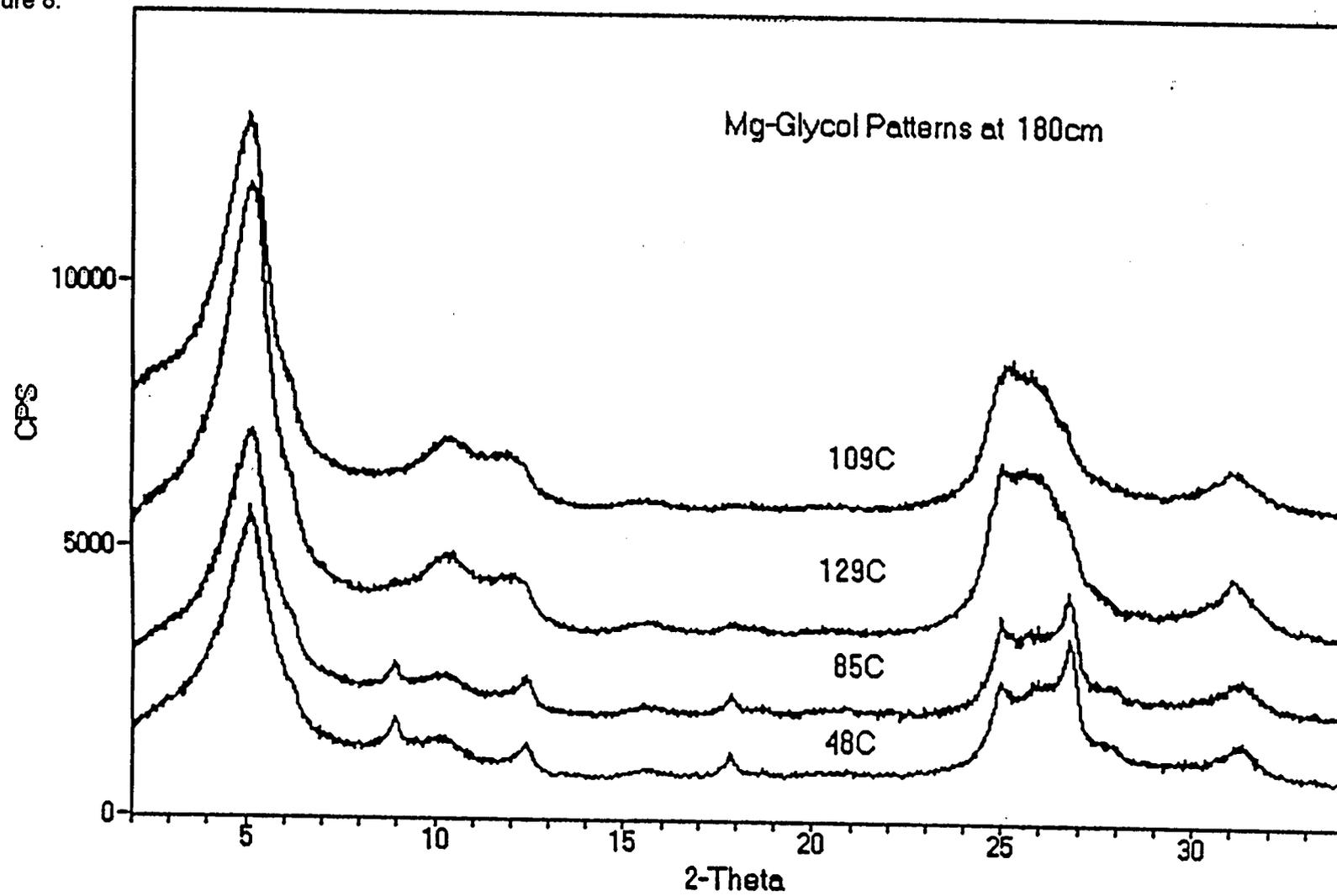


Figure 8. XRD patterns of soils in SE to NW transect at 180 cm

Figure 8.



Discussion

The mantle of black clay revealed in figures 2, 3, and 4 indicates that this material has been continually augmented by the alluvial source streams. The streams flowing from the foothills carry sediment to the northwest of the wetland where the low gradient causes the water to slow down resulting in accumulations of sediment. The clay deposit becomes thinner toward the southeast corner of the wetland furthest from the sediment source indicating that the clay has been transported from the foothills to the wetland via Jackson and Frazier creeks. The presence of smectite in the XRD patterns is consistent with the observed physical characteristics of the surface soil. The soil is very sticky and very plastic and one can form very long plastic ribbons. The soil typically exhibits extremely slow hydraulic conductivity and ponds water most of the year. The water holding capacity and shrinking and swelling nature of the soil correspond well with the observed mineralogy. The near permanent saturation and mild climate of the wetland facilitate both the transformation of alluvial source material to smectite and the maintenance of the smectite in its present form once deposited. The kaolinite identified in the samples is most likely a residual source material from the surrounding foothills, while the vermiculite may be an authigenic mineral from the wetland or a detrital product of the foothill alluvium.

The olive brown silty clay layer underlying the black clay also seems to be alluvial as the contour of the deposit parallels the topographic surface. This deposit is similar in mineralogy to the surface clay, indicating a common source. The clay does have significantly less organic material, indicating that the environment of deposition was quite different than that observed at the present. The consistency and

texture of the olive brown silty clay is similar to the Malpass member of the Willamette Formation. It also lies in a position that is consistent with Malpass deposition in other locations in the Valley. However, mineralogical analysis is inconclusive regarding whether it is the malpass unit and there may have been a great deal of mixing between the olive brown silty clay and the underlying sediment obscuring the nature of the clay. It seems certain that the olive brown silty clay is a distinct deposit that differs from both the upper alluvial deposit and the underlying Irish Bend silt, but its origin is uncertain given the scope of the present study.

The configuration of the light olive brown silty clay loam deposit suggests that its origin is not alluvial. The silts do not follow the contour of the hillslope, but appear to have been scoured away by erosion due to the smooth, level boundary between the olive brown silty clay and the Irish Bend silt. The surface of the Irish Bend silt may also have been reworked with the overlying olive brown silty clay.

The clay mineral assemblage from the silts displayed in the diffraction patterns at 48C and 85C is clearly distinct from that of the overlying deposits. Glasmann and Kling (1980) identified the mineral assemblage shown in these patterns as Irish Bend silt. They contended that the presence of clay size mica indicates that the silts were formed from a source other than the local foothills. The presence of this deposit under the wetland soils indicates a clear stratigraphic boundary underlying the alluvial deposits in the southern part of the wetland. The silts have been reported to extend to 80 meters elevation (Glasmann, et al., 1980) therefore leaving the distinct possibility that the silts are present at the higher elevations, but lie lower than 180 cm at these points.

Conclusions

The absence of convincing evidence for the presence of the Malpass and Greenback stratigraphic deposits suggests that either these deposits were never located in this area, or that they have been scoured away by erosional forces. It would be difficult to ascertain if the previous overlying stratigraphic deposits were present or not at higher elevations without intensive sampling of higher surfaces surrounding the wetland. However, assuming that these deposits were laid down over the Irish Bend deposit, it is possible that a meander channel of the Willamette River scoured a path through the wetland from the direction of Stewart Slough toward the north following a path back to the main stem of the Willamette via Frazier Creek ditch. This scouring would have removed the later Pleistocene deposits as well as any older alluvium leaving only the Irish Bend silt deposit exposed. Recent alluviation then may have covered the of the wetland with locally derived clayey sediments obscuring and perhaps mixing with the upper part of the underlying Irish Bend deposit.

Chapter III. Hydrology of the Jackson-Frazier Wetland

Introduction

The driving force for the creation and maintenance of wetlands is water. Water moves over and through wetland soils acting as a solute for the transport of inorganic ions and as a catalyst in biochemical and geochemical transformations. Certain aspects of soil genesis, namely, the addition, removal, transformation and translocation of soil materials are facilitated by water (Daniels and Hammer, 1992). Water influences soil genesis directly through both the direction of water flow and the duration of soil saturation. Water flows vertically through highly permeable soil but moves laterally through soils with restrictive layers of slow permeability. In wetland areas, layers of slow permeability create situations in which water saturates surface horizons for long periods of time and may collect on the surface of soils. Long-term saturation and ponding of water result in quite different soil chemical and physical transformations than those found in highly permeable soils with unimpeded drainage.

The dynamic nature of water movement in a wetland and the lack of long term data for most wet soils has rendered hydrologic interpretations as the least useful parameter in wetland identification (Environmental Laboratory, 1987; Federal Interagency Committee, 1989). Long term hydrologic monitoring provides the most reliable measure of the depth and duration of saturation of a site, but it is difficult and expensive research. In order to make more accurate assessments of the processes involved in soil genesis, detailed hydrologic information is needed.

Problem Statement

The defining concept in identifying, and later delineating, regulatory wetland boundaries has been "wetness". Areas that are saturated with water near the surface for prolonged periods of time exhibit distinct vegetation and groundwater conditions. Although the slowly permeable layers of wetland soils promote near surface saturation, wet soils do not always have uniformly saturated horizons, nor is the pattern of moisture in these soils always similar.

Regulatory wetland delineations currently focus on three factors: hydrology, vegetation, and soils. According to these factors, a wetland soil is defined as one that is saturated to the extent that it will support hydrophytic vegetation (Cowardin et al., 1979). However, the degree of saturation in this definition is broadly interpreted, and soil is often assumed to be uniformly saturated. Soils in the "aquic" moisture regime are saturated from ground water or the capillary fringe. These two mechanisms of saturation are distinct and result in different conditions of available water for plants. In the past, the aquic moisture regime designation in soil taxonomy did not reflect these specific field conditions of saturation (Dudal, 1990). Revisions of the definition of the aquic moisture regime have resulted in more descriptive possibilities for soils that exhibit surface or groundwater saturation (Bouma, 1990; Soil Survey Staff, 1992).

Identifying wetland areas based primarily on the presence of hydrophytic vegetation is also problematic due to many diverse plant adaptations to wet and dry environments (Tiner, 1991). A more specific classification of wetland areas using detailed soil hydrologic information would help link soil saturation characteristics with

plant community distribution and clarify the classification of wetland areas and wetland soils.

Groundwater saturation can occur either near the surface as a perched water table, or as the complete saturation of an entire soil profile. Two types of saturation in soil are defined in Keys to Soil Taxonomy, episaturation and endosaturation (Soil Survey Staff, 1992) First, saturation of the entire profile can be caused by water moving upward from a deeper apparent water table. This type of saturation is defined as endosaturation and is typical of medium and coarse textured soils that have continuous saturation throughout all of the horizons if the water table rises high enough. Episaturation results from the saturation of surface layers of soil due to horizons with slow conductivities. These soils are saturated at or above the surface but are unsaturated in one or more subsurface horizons. Fine textured soils often display surface saturation, or episaturation, due to the ponding of water in areas characterized by slowly permeable soil layers near the surface.

The type of soil saturation is important for the proper classification of soils and also provides information for the interpretation of the chemical transformations taking place within the soil profile. In addition, hydrologic data can characterize the nature of water flow through large areas of soil, as well as water flow within particular soil pedons.

Piezometer Theory

Long term observations of the water in the subsurface must be made to describe the type of saturation in a soil. This information can be efficiently obtained by using permanently installed monitoring tubes known as piezometers. The level of water in piezometers corresponds to hydraulic head (Freeze and Cherry, 1979).

Hydraulic head is an indication of the energy status of water in water flow systems. The energy status is often referred to as the total potential of a groundwater system. The energy status of water standing at some point in an open pipe, such as a piezometer, above some reference elevation, or datum, can be described by the Bernoulli equation (Freeze and Cherry, 1979) which describes the energy status of a flowing liquid in terms of kinetic, potential, and pressure energies:

$$h = \frac{v^2}{2g} + \frac{p}{w} + z$$

Where,

- h-total potential
- v-velocity
- g-acceleration of gravity
- p-pressure
- w-specific weight of the fluid
- z-elevation potential

In soils, the velocity term ($\frac{v^2}{2g}$) can be disregarded due to the slow movement of water through the soil. The resulting equation, which describes the level of water found in the piezometer tube, consists of the sum of the pressure potential and the elevation potential:

$$h = \frac{p}{w} + z$$

The flow of water responds to a potential gradient in the subsurface environment where water flows from regions of high total potential to areas of low total potential. Consequently, the levels in several piezometer tubes can be interpreted as the contour of the water surface, or piezometric surface.

Measuring piezometric surfaces at varying depths in the soil profile is useful in determining pressure relationships of water moving in the subsurface. The pressure relationships are especially important in situations where restrictive layers of soil confine the water within the upper or lower soil horizons. The piezometric surface does not necessarily correspond to the position of the free water surface in the soil. A water table is recognized by specific characteristics of pressure. The pressure of the water above a water table is lower than atmospheric pressure, while the pressure of water beneath the water table is greater than atmospheric. The theoretical definition of a water table itself is the underground water surface at which the pressure is exactly equal to atmospheric pressure (Domenico and Schwartz, 1990). A piezometric surface indicates the level to which water will rise in a tube in response to the total potential of the hydraulic system.

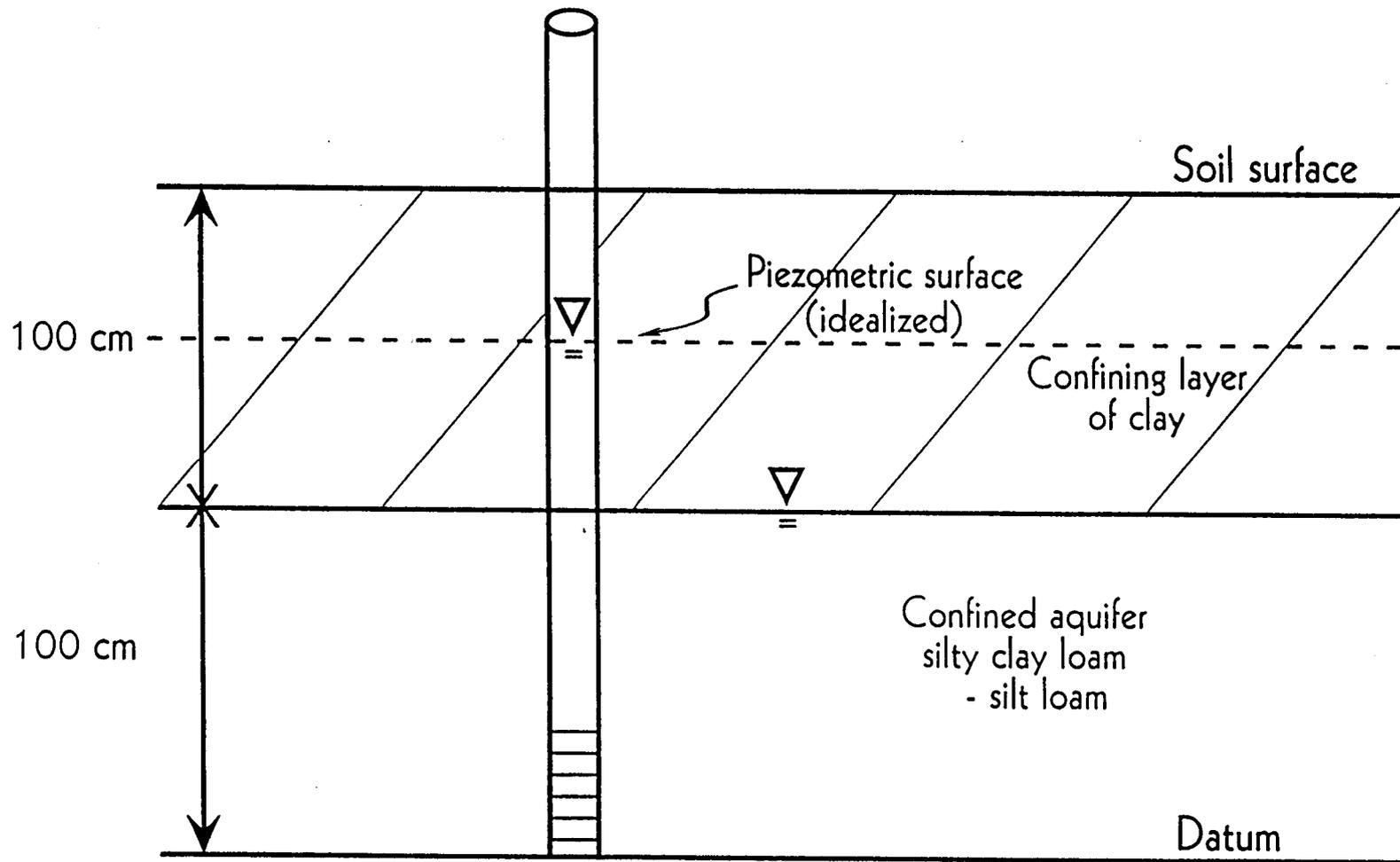
Aquifers

Confined and unconfined aquifers control the behavior of groundwater movement in soil (Freeze and Cherry, 1979; Domenico and Schwartz, 1990). Determining the type of aquifer present is important for the proper interpretation of observations of the piezometric surface.

In confined aquifers, the soil is saturated up to the lowest extent of a confining layer. Under conditions of confinement, the total potential of the water is greater than the water located in the confining zone. A piezometer which extends through a confining soil layer and is open in the confined aquifer will have a piezometric surface which is higher than the surfaces of piezometers which are open only in the confining zone (fig. 9).

Figure 9. Piezometric surface and the confined aquifer

Figure 9.



In unconfined aquifers, the water moves freely through all soil horizons. The total potential at each point in the soil is equal at all depths (fig. 10).

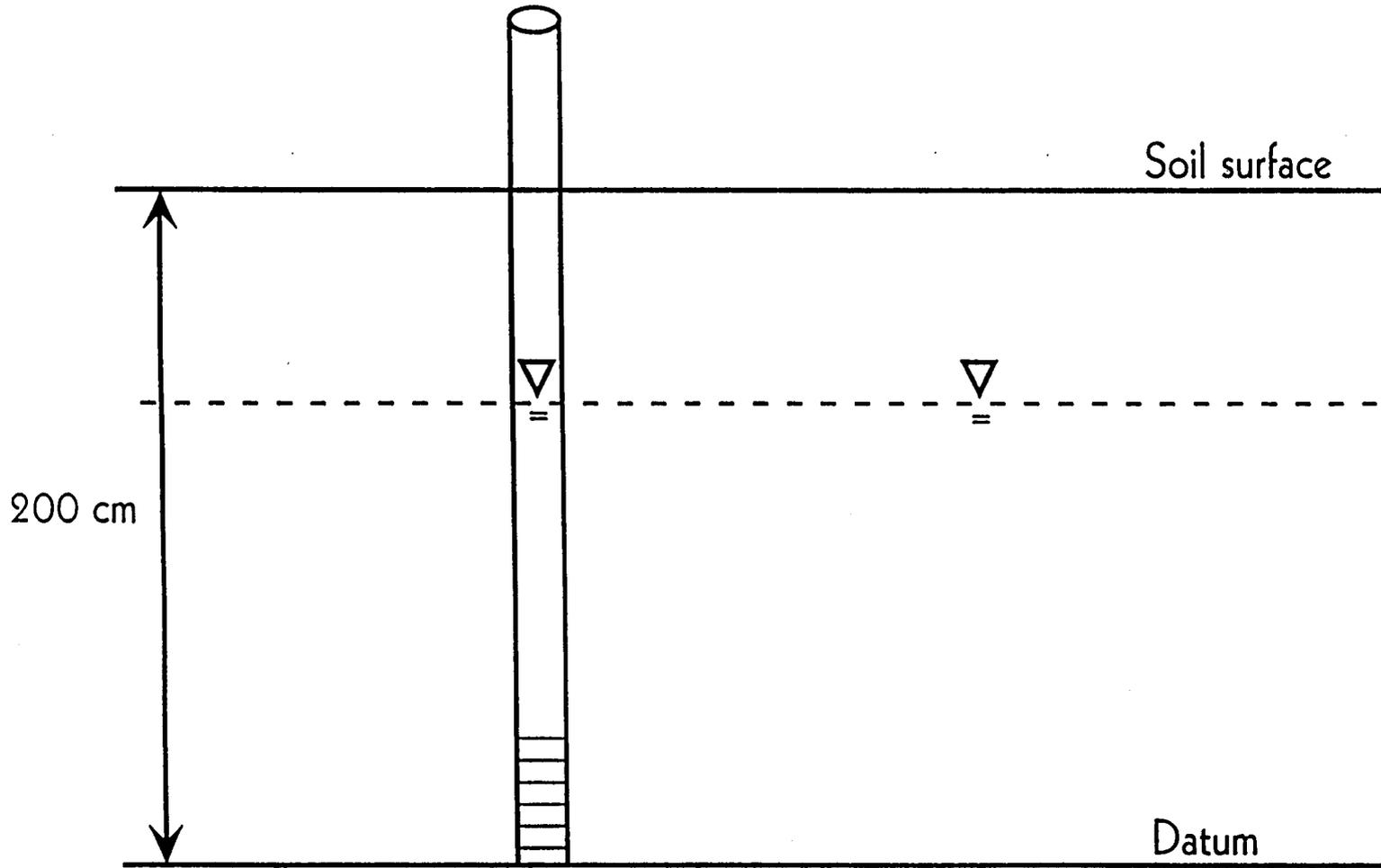
Methods

Piezometers were used in the present study to observe the changes in the piezometric surface over time in a soil profile. The piezometer wells were designed to measure the level of "free" water in the soils and the response of this water to the pressure head exerted at each depth. Piezometers were constructed from 3/4 inch (1.9cm) o.d. schedule 200 PVC pipe. Seven horizontal slots 1 cm apart were cut in the bottom of each pipe. The slotted end of the PVC pipe was then covered with textile fabric to prevent clogging of the tube. An acrylic tube containing a styrofoam float was placed inside the PVC pipe to register the level of water in the tube. The piezometers were placed in the soil by boring a 2 inch (7 cm) hole to the required depth in the soil. The holes were then excavated an additional 1 inch (2.54 cm) and the bottom filled with fine sand. The piezometer tube was then placed in the hole and the space between piezometer and soil backfilled with fine sand to the top of the perforation of the PVC pipe. The hole was then backfilled with bentonite to seal the tube from surface leakage. Piezometer wells were placed in triplicate in the monitoring plots at depths of 25, 50 and 100 cm. One 200cm piezometer was placed in each plot. Readings were taken weekly by removing the acrylic tube and measuring the depth from the soil surface to the styrofoam float. A more detailed description of piezometer construction and installation is provided by Austin (1994).

Data for the piezometers at plots 1, 2, and 3 are displayed in figures 11, 12, and 13 respectively. Data collection began in late October, 1992 and was terminated for this project in early March, 1994. The data reflect a cycle from an early wetting

Figure 10. Unconfined aquifer

Figure 10.



period through a saturated period from December to May, then a drying trend from June to October, and finally another period of wetting and saturation from November to March.

Three plots were chosen for the study within the wetland. Each site was located on soils with similar characteristics, but had different vegetation cover. The scope of the data presented does not include an evaluation of soil-plant interrelationships, however, a description of the plants at each site and some general observations are offered here for reference in future studies.

Plot 1 was placed in an area of water parsley (*oenanthe sarmentosa*). Plot 2 was in a slough sedge (*carex obnupta*) meadow, and plot 3 was placed within reed canary grass (*phalaris arundinacea*). Plot 1 consistently had the most water on the surface indicating that the water parsley may tolerate large amounts of water. Plot 3 was the driest area which may have been caused by evapotranspiration by the reed canary grass. Surface water did not accumulate during heavy precipitation in the spring when evapotranspiration was high. Plot 2 was intermediate in surface inundation between the very wet condition at plot 1 and the drier surface at plot 3.

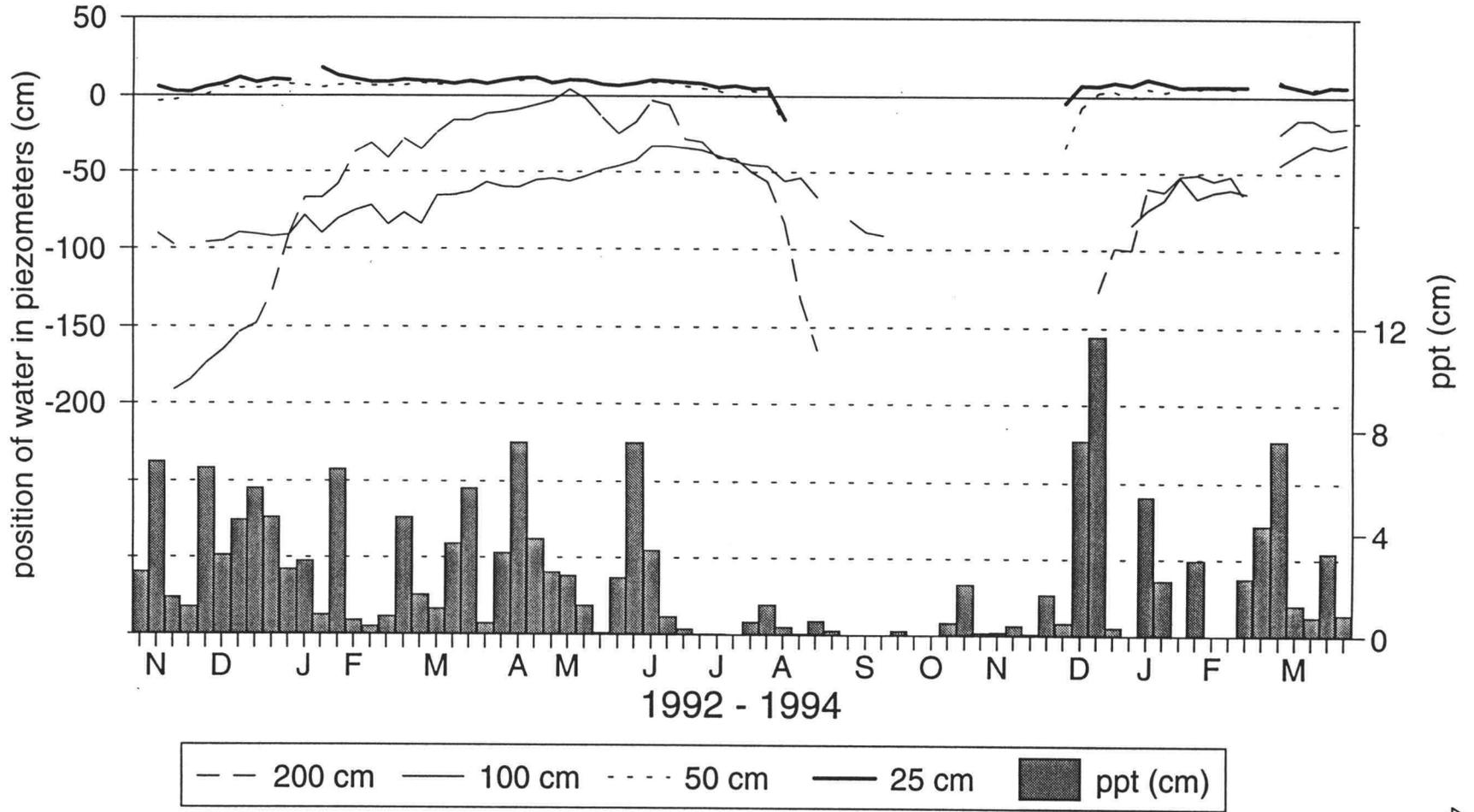
Results and Discussion

Plot 1

Water levels above the soil surface, or 0 in fig. 11, were present in the 25 and 50 cm piezometers from November 1992 to late July 1993, indicating surface saturation throughout this period (fig. 11). Water levels began to fall in early August 1993 at both depths, and no water was present in piezometers from late August 1993 through the middle of November 1993. Water levels rose again in late November

Figure 11. Precipitation (bottom) and position of water (top) as observed in piezometers at depths of 200 cm, 100 cm, 50 cm, and 25 cm at the Bashaw plot 1 at the Jackson-Frazier wetland in Benton County. The 0 cm position is the location of the soil surface.

Figure 11.



1993 and were above the soil surface by early December 1993. These water levels were maintained until the end of data collection in March 1994. It is interesting to note that the water level in the 25 cm piezometer was higher than that in the 50 cm piezometer upon initial wetting in November 1992 and 1993, indicating that the upper part of the soil became saturated from the top down. Piezometric water levels above the soil surface at 25 and 50 cm depths indicate that the soil is saturated from the soil surface to a depth of 50 cm for most of the year.

The water levels in the 100 cm piezometers were close to 100 cm below the surface in early November, indicating that initial wetting at this depth was slower than at the 50 and 25 cm depths. The water levels slowly rose to approximately 40 cm below the soil surface in early June 1993 and then dropped off until no water was noted in early September 1993. Water was observed again in the piezometers at 75 cm below the surface in December 1993. The water levels then rose steadily from December 1993 to March 1994 when data collection terminated.

The fact that water levels in the 100 cm piezometers were deeper than 50 cm for a long period of time indicates that water is not moving downward through the soil very quickly. Water that flows into the wetland from Jackson and Frazier creeks moves across the surface slowly due to the gentle gradient of the wetland toward its outlet points. The extremely slow permeabilities of the surface soil minimize the vertical infiltration of water, which maintains the surface ponding of the soils. The water levels in the piezometers remains at 6 to 10 cm above the soil surface for most of the year, corresponding to the level of water ponded on the surface of the soil. Surface runoff exits the wetland via Stewart Slough and Frazier Creek ditch .

Buffkin (1985) conducted a study to assess water inflow, outflow and storage in Jackson-Frazier during 1984-85. This study showed that the peak flow lag time between the input at Jackson-Frazier creek and output at Stewart slough was longer at the onset of the rainy season in November than later in the rainy season in February, indicating that water was moving across the wetland more rapidly in February. This decrease in lag time indicates that groundwater detention storage decreased after the soils were completely saturated. After the soil becomes completely saturated, the detention storage of the wetland decreases and surface runoff is the dominant flow pathway for incoming water.

The piezometers at the 100 cm depth indicate that the subsurface (50-100 cm) was unsaturated until late winter/early spring, corresponding with the latter part of the rainy season. This observation is in good agreement with Buffkin. Once the subsurface became saturated, the storage capacity of the wetland was very limited.

The water in the 200 cm piezometer was close to 200 cm below the surface in early November 1992 and climbed rapidly through late January 1993 to a level of 40 cm below the soil surface in response to abundant rainfall (fig. 11). The water level alternated between falling and rising in response to sporadic rainfall throughout February 1993, then rose steadily in March and April 1993, reaching a peak in May 1993 for a brief time. The level then dropped until late May after which the water level rose briefly in response to a heavy rainfall. The water then dropped off quickly until no water was detected in the piezometer in late August 1993. The water level in the 200 cm piezometer was apparent again in mid December 1993 after a period of intense rainfall. The level climbed steadily until early January 1994 where it leveled

off due to sporadic rainfall. A slight drop at the end of a dry period was followed by a sharp increase with heavy rain in late February 1994.

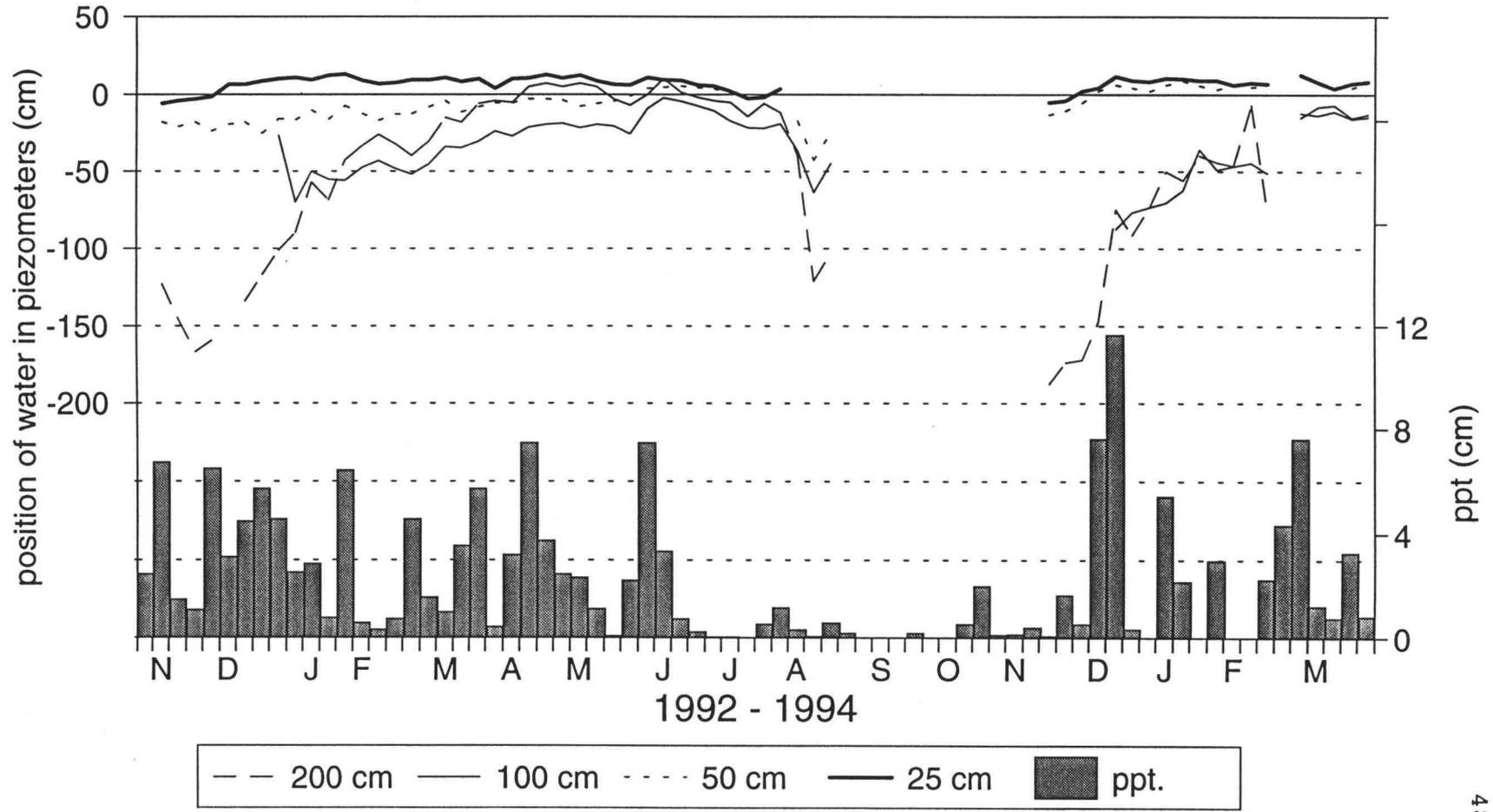
The piezometers at 100 cm depth were positioned at the transition zone between slowly permeable clay and more permeable silty clay material that underlies the southeast portion of the wetland (chapter II). The slowly permeable black clay described in chapter II forms an aquitard above the silty clay and the Irish Bend silt. The piezometers positioned at 200 cm depth are open beneath the aquitard and are located in the Irish Bend silt that underlies the black clay and silty clay. Water was present in the 100 cm piezometer before the water level in the 200 cm piezometer indicated that water was present at a level of 100 cm in the 200 cm piezometer. Therefore, the water must have moved downward from the surface to the 100 cm piezometers before the soils swelled completely. During the early portion of the wet season, the water levels in the 100 cm piezometers rose slowly (figs. 11 and 12) in response to the slow infiltration of surface water downward through macropores in the black clay. Once the pores located in the aquitard swelled shut due to the shrink-swell nature of the clay, the pressure of water rising from the silts dominated the water regime at 100 cm.

Plot 2

The water levels at 25 and 50 cm at plot 2 were very similar to those at plot 1 (fig. 12). The 25 cm water levels rose to levels above the soil surface by the end of November 1993 indicating surface saturation. The water remained above the surface until late July 1993 after which there was a rapid drying and no indication of water in the piezometers until November 1993. The water levels rose again from mid November through March 1994 at 25 cm.

Figure 12. Precipitation (bottom) and position of water (top) as observed in piezometers at depths of 200 cm, 100 cm, 50 cm, and 25 cm at the Bashaw plot 2 at the Jackson-Frazier wetland in Benton County. The 0 cm position is the location of the soil surface.

Figure 12.



The 50 cm water level moved from 20 cm below the surface upward from November 1993 to April 1993 where it finally indicated a position near the soil surface. After a brief fall in May 1993, the level climbed to above the soil surface through late May, June and early July 1993. The water level then fell rapidly in late July and August 1993. There was a brief upward movement of water in the 50 cm piezometer in August 1993 after which drying was rapid. The water levels remained below 50 cm until November 1993 when the water moved upward again. The water was above the soil surface from late November 1993 through March 1994.

The 100 cm piezometers indicated water levels below 100 cm from November through mid December 1992. An initial water level measurement showed the total potential at 25 cm, but it quickly fell to 60 cm in late December 1992. The water levels began rising at this time and climbed to the soil surface in late May 1993. The levels then fell until late August 1993 when there was a brief increase in total potential and then complete drying. The levels began rising above 100 cm in early December 1993 and climbed steadily until February when a large jump in water level was recorded due to a significant rainfall in late February.

The water level in the 200 cm piezometer was initially observed at 125 cm below the soil surface in early November 1992, but then fell to 160 cm by mid November 1992. The level climbed steadily to 60 cm in January 1993 when it alternately fell and rose from January 1993 until April 1993 where it attained a level above the soil surface. There was a brief decline in water level in late May 1993 and then a surge above the soil surface in early June 1993. The decline in water level was rapid after June 1993 with only a brief rise during the first week of August 1993. The water level was next indicated above 200 cm in November and rose quickly

throughout November and early December 1993. The water level rose rapidly during the first week of February 1994, dropped down during the following week, and spiked upward to 20 cm below the surface after the large rainfall in late February 1994.

Plot 3

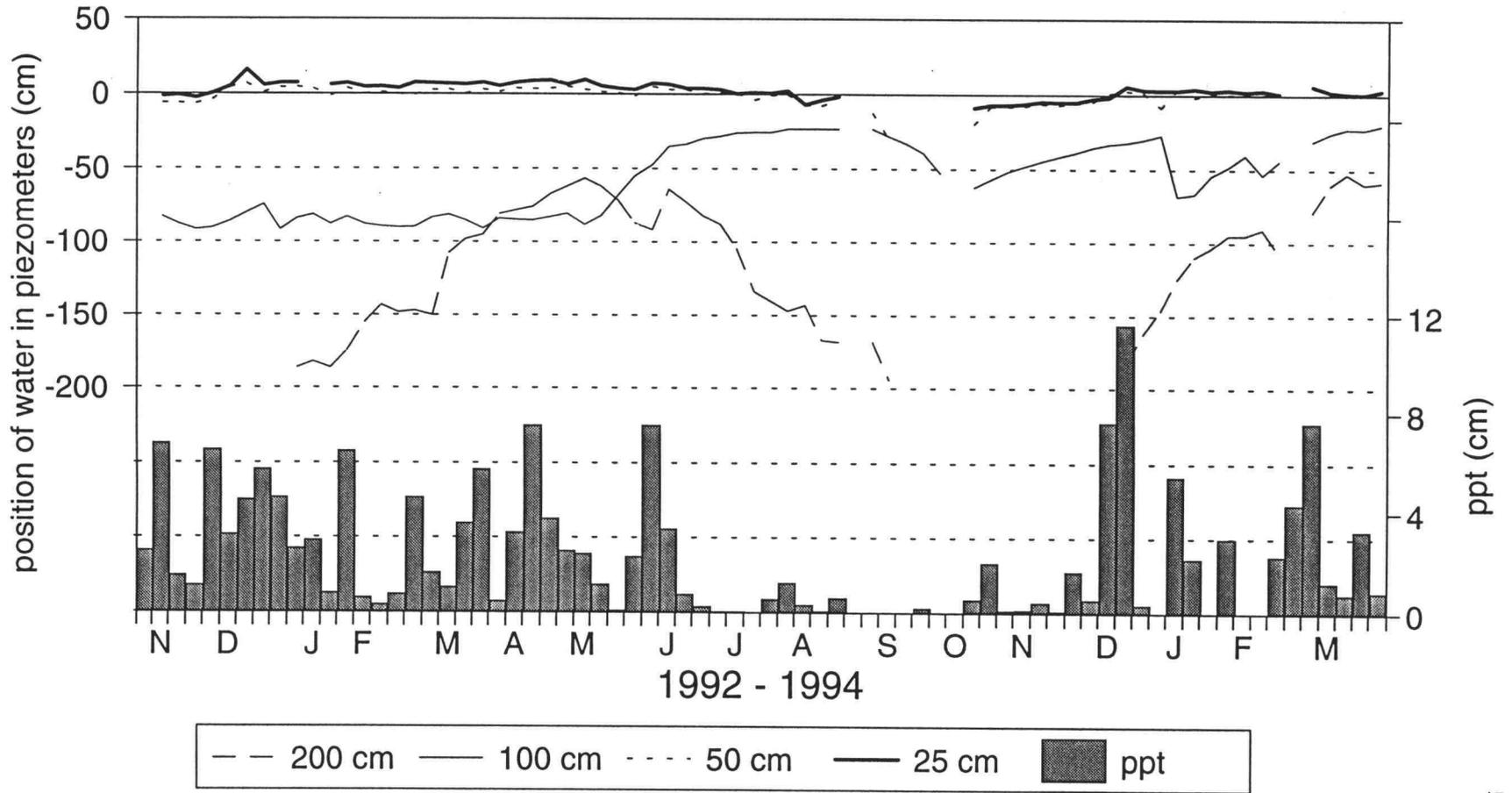
The 25 cm piezometers indicated surface saturation from November 1992 through late July 1993 (fig. 13). The water level then dropped, rose again to near positive, and then rapidly dropped in mid August 1993. Water levels were again observed at about 15 cm below the surface in October 1993. The water levels then rose above the soil surface through October and November 1993 and remained there from December 1993 through March 1994.

The water level in the 50 cm piezometer was near the soil surface in November 1992 and then rose to the surface in December 1992. The water level position above the surface was maintained until late June when the levels began to drop. The levels dropped below 50 cm in September 1993, and returned to 25 cm below the surface in early October. The water level then rose to above the soil surface from October to December 1993 and was maintained through March 1994.

The water levels in the 100 cm piezometer were near 100 cm below the soil surface in November 1992. The levels fluctuated a bit, but did not increase until early May 1993. The water then rose to 25 cm below the surface in late May 1993 and was maintained at that point until early September 1993. The levels then decreased in September 1993, but never fell further than 60 cm below the soil surface. The water level began rising again in October 1993 and rose until late December 1993 when it dropped to 60 cm below the surface. The level then rose steadily to 25 cm below the surface in March 1994.

Figure 13. Precipitation (bottom) and position of water (top) as observed in piezometers at depths of 200 cm, 100 cm, 50 cm, and 25 cm at the Cove plot 3 at the Jackson-Frazier wetland in Benton County. The 0 cm position is the location of the soil surface.

Figure 13.



The water level in the 200 cm piezometer was deeper than 200 cm below the soil surface until late December 1992. The water level then rose steadily through January and leveled off at 150 cm below the surface in February 1993. The water level then rose quickly through March and April 1993, and peaked at 60 cm below the surface in May 1993. The level then fell through May 1993, surged upward in early June 1993 and then fell steadily until moving below 200 cm in September 1993. The water level in the 200 cm piezometer was observed again at the end of December after the rain began to fall. The water level rose steadily until early February 1994 when it declined during a dry period. The water level then rose to 50 cm in late February, 1994, and leveled off.

Conclusions

The correspondance of the surge in water levels in the 100cm and 200 cm piezometers indicates that water flowing under increased pressure in the Irish Bend silt is influencing the water levels in the piezometers at plots 1 and 2. However, at plot 3, the black clay aquitard extends beyond two meters and the influence of the water under pressure in the Irish Bend silt is not apparent. The water level in the 200 cm piezometer at plot 3 rose above the 100 cm head level only briefly during April 1993 and not at all during the period of observation in 1994. An increase in water levels in the 100 cm piezometers corresponding to heavy precipitation occurred several times at plots 1 and 2, but most notably during the fourth week of December 1992, the first week of February 1993, and the third week of March 1993.

The 200cm piezometers at plots 1 and 2 measured water levels from water moving under pressure in the Irish Bend silt below the black clay aquitard. The movement of water through the silts was indicated by the water level in the 200 cm

piezometer above the water level of the 100 cm piezometers. If infiltration was the dominant recharge pathway and the aquifer was unconfined, then the water levels would have been at the same position in both the 200 cm and 100 cm piezometers. The observations show a higher pressure potential in the 200 cm piezometer, indicating water moving in a confined aquifer. This phenomenon has been noted in similar soils in the Willamette Valley (Boersma and Simonson, 1972) and may be a common feature due to the stratigraphy of the valley.

Differences in the water level measurements among the plots can be accounted for by different layers of permeability. Plots 1 and 2 have a contact with the Irish Bend silt at 135 cm and 120 cm respectively. Water is detected above 200 cm in the 200 cm piezometers soon after the first rains. The water at 200 cm is primarily due to the flow of water horizontally through the Irish Bend silt. If the water flowed through cracks, the water level measurements would taper off and surface ponding would not be as prevalent. Plot 3, which does not have a contact with the Irish Bend silt above 200 cm, showed a significant lag between the initial precipitation and the water level above 200 cm. The water level in the 200 cm piezometer never rises very high, indicating a lower pressure potential in the piezometer. The movement of water is more restricted through the thicker aquitard at plot 3. Plots 1 and 2 show water levels which rose to the surface and above. These results correspond well with the map of deposits in chapter II. The thicker deposits near the confluence of the creeks restricts the subsurface movement of water upward and downward in the profile.

The saturation at all levels is related to precipitation patterns. The response at 25 and 50 cm is very rapid with surface saturation occurring very soon after the

first rains of the season. Water flowing from Jackson and Frazier creeks off the watersheds moves over the surface and rapidly saturates the soil at shallow depths. Once the surface soil is saturated, the very slow permeabilities of the soil prevents water from infiltrating quickly, and a prolonged period of surface saturation begins because of excess surface water. The surface saturation is perpetuated by rainfall from November to July. Only after prolonged drying in July and August does the water begin to disappear. Soils saturated with water primarily from the surface, and not in contact with a lower aquifer are classified as episaturated (Soil Survey Staff, 1992).

The slowly permeable montmorillonitic clay creates saturation of the soil to a depth of 50 cm which results in the accumulation of water above the soil surface. The clay also prevents the infiltration of water downward in the profile leaving a relatively unsaturated zone between 50 and 100 cm for several months.

Two water tables are present throughout the rainy season in these soils. The water rising from the lower water table under pressure eventually causes the water to flow into the clay zone. The saturation of the clay layer prevents additional detention storage of the wetland and increases surface runoff. The presence of the confined aquifer influences the subsurface movement of water. The low vertical permeability of the clay impedes infiltration, but some of the water movement would be vertical if the lower horizons were unsaturated. The presence of the confined aquifer creates horizontal flow below the soil surface by forcing water to move along the topographic gradient of the Irish Bend silt.

It may seem obvious to define the conditions as episaturation because of the influence of the impermeable layer and the ponding of surface water. However,

wetlands which have significant sources of water from below might have to be considered special cases of endosaturation. The water pressure from the subsurface influences the surface processes such as ponding, detention storage and runoff. The wetland is not episaturated for the entire year, but is endosaturated during the late winter and early spring. Saturation at this time corresponds to the period of most intense reduction in the soil and has a strong influence on the transport and distribution of altered soil constituents such as iron and manganese (chapter IV).

Chapter IV. Redoximorphic Character of the Jackson-Frazier Wetland

Introduction

The chemical reactivity of a soil can be correlated with hydrologic measurements to determine how soil saturation affects microbial action in the soil. The activity of microbes that catalyze chemical reactions in soil is sensitive to soil aeration. Oxidation states of inorganic and organic compounds in soil will change as the soil becomes wet and can be used as general indicators of soil saturation.

Oxidation and reduction state (redox) of a soil is linked to biogeochemical reactions under either anaerobic or aerobic conditions. Organic carbon is oxidized by aerobic microorganisms in soils when there is an ample supply of oxygen. As water fills soil pores, air is displaced, the soil becomes depleted of oxygen, and the replacement of oxygen by diffusion through water filled pores is extremely slow. Under saturated conditions aerobic respiration ceases, anaerobic organisms become the primary decomposers of organic matter, and compounds other than oxygen must be utilized as receptors for electrons supplied by anaerobic organic matter decomposition.

Soil classification in the United States (Soil Survey Staff, 1992) recognizes the importance of these processes in soil development through the aquic conditions diagnostic feature. Aquic conditions imply that the soil becomes anaerobic some time during the year, and requires confirmation of three specific factors: saturation, reduction, and redoximorphic features (Vepraskas, 1992). Saturation must be confirmed by observation of water tables, and reduction must be confirmed either with measurements of redox potential or by a positive reaction with alpha, alpha-dipyridyl. Redoximorphic features are used as field indicators of anaerobic conditions in soils.

Saturation, anaerobiosis, and chemical reduction result in distinct soil colors because of the transformations of iron and manganese. Redoximorphic features consist of color patterns that indicate areas of depletion of iron and manganese, concentrations of oxides of iron and manganese, and depletions and translocations of clay. A complete guide to the identification and description of redoximorphic features has been prepared by Vepraskas (1992).

Field measurements of present conditions are used to verify whether reducing conditions are currently present in a soil and if the redoximorphic features represent past conditions or are currently forming. The correct interpretation of redoximorphic features can only be completed after soil properties have been measured to verify the type of feature associated with the duration of saturation and the intensity of reduction.

Problem Statement

Soil morphology has long been used as a tool to interpret saturation and reduction in soils (Soil Survey Staff, 1975). Soil color was the primary indicator of wet conditions in soil morphological descriptions using keys to soil taxonomy prior to 1992. However, color descriptions used to identify soil saturation were limited to low chroma mottles (chroma ≤ 2 , value ≥ 4) and high chroma mottles. The color of low chroma mottles was interpreted as the color of uncoated primary mineral grains after the stripping of iron from the silt and sand surfaces due to reduction and translocation. Brown, strong brown, and reddish yellow high chroma mottles were identified as precipitated iron oxides such as goethite and hematite (Schwertmann and Taylor, 1989) indicating alternating wet and dry periods. Techniques for making detailed descriptions of redox features were not available in previous keys to soil

taxonomy. There was no distinction between the effects of surface ponding and the saturation of the entire profile. Surface saturation due to perched water tables at shallow depths may result in different soil morphological distinctions than saturation of an entire profile. Surface water gley soils (Fanning and Fanning, 1989) represent an important realm of soils which merits more intensive study and better classification in order to guide the better use and management of areas which contain these soils (Brinkman, 1979; Schwertmann and Fanning, 1976).

Due to the limitations of descriptive techniques, several problems in interpreting reduction in soils arose during recent research. Veneman et al. (1976) observed that in some poorly drained, fine textured soils in Wisconsin, reducing conditions still persist within soil peds, even after the water table has receded. Microsite variation in soils that retain water results in diverse anaerobic and aerobic environments. Pickering and Veneman (1984) found that wet, but unsaturated conditions can significantly affect soil morphology through the translocation of reduced iron. They found that the upward movement of reduced iron and its subsequent precipitation led to chromas of six or higher. Couto et al. (1985) found soils with low organic matter or organic matter which is not easily decomposed will not show reduction, even when saturated. The lack of microbial activity due to limited organic matter points to the essential role that micro-organisms play in the cycle of electron availability in soils. The micro-organisms must be actively using organic matter as an energy source for electrons to become available for the reduction of inorganic and organic compounds. Daniels et al. (1973) found that abundant oxygen limited the reduction of iron in some North Carolina aquults and adults. The highly aerated soil water inhibited reduction even though the soil was saturated. Soils that

contain no iron will show no visible signs of reduction. In soils that contain no iron, there may be long term saturation but no low chroma mottles visible in the soil profile. High chroma mottles will not appear during periods of alternating wet and dry cycles in low iron soils as well (Franzmeier et al. 1983; Vepraskas and Wilding 1983).

Other processes related to reduction in soils are significant in interpreting a soil's morphologic signature and assessing its redoximorphic character. Brinkman (1979) described the process of "ferrolysis", the acidic dissolution of clay by the formation of hydrogen ions through the alternating reduction and oxidation of iron. Vepraskas and Guertal (1992) discuss the translocation and accumulation of Fe, Mn and clay through soil macropores as a diagnostic feature of redoximorphic conditions.

Revisions to the definition of the aquic soil moisture regime have attempted to expand the definitions for redox processes in soil. The biogeochemical process of reduction and its association with soil saturation is now described by citing all pertinent redoximorphic features found in a soil. Redoximorphic feature description is an attempt to identify both color patterns and solid forms found in soil that are the result of redox processes and distinguish them from other pedogenic formation factors.

However, the application of redoximorphic feature description is limited by the knowledge of existing conditions in soils and the correlation of measured parameters to observed redoximorphic features. Redoximorphic features must be validated locally through long term monitoring of selected soils for the descriptions to be applied to other similar soils in an area during field soil descriptions.

Theory of Reduction/Oxidation

Soils are inherently metastable. That is, the soil is in a state of temporary stability between the complete oxidation of soil organic carbon to CO₂ by microorganisms and no respiration at all. Equilibrium thermodynamics provides us with a set of rules governing the reactions of carbon in soils. Analysis of the oxidation of organic carbon to CO₂ in the presence of oxygen suggests that CO₂ is the stable species and the thermodynamic reaction should go to completion. However, this process may be mediated by the transport of oxygen to the site of aerobic respiration, the variability of microbial populations that control thermodynamic reactions, and the annual reduction of carbon via photosynthesis (Bartlett, 1986).

Nonetheless, thermodynamic relationships are used to simplify our understanding of chemical reactions in soil. The use of thermodynamic stability is of great importance in the description of oxidation and reduction (redox) reactions in soils.

Oxidation and reduction in soils is completed by the transfer of electrons from organic carbon to an electron acceptor, such as oxygen. The species that donates an electron is oxidized, that which accepts an electron is reduced. Redox reactions are measured in soil with a platinum electrode/reference electrode pair (Cogger and Kennedy, 1992; Szogi and Hudnall, 1990; Faulkner et al., 1989). The platinum electrode is an inert means of transferring electrons in the soil. An ideal model for describing an electrode system is described by Drever (1988). In this representation, the platinum system consists of two half-cells connected by a salt-bridge. The platinum electrode is immersed in one of the half-cells and electrons flow through it to the other half-cell. The flow of electrons from the platinum cell to the adjacent cell

creates a potential which can be measured with a high impedance voltmeter. The cell is now considered the soil and the circuit from the soil through the platinum electrode is completed with a standard reference electrode connected to the soil with a KCl salt bridge (Veneman and Pickering, 1983).

Theoretically, values for the potential in soil can be calculated using the Nernst equation and half reactions involving electron donors and acceptors (Drever, 1988; Sposito, 1989). The calculation of a potential through the transfer of an electron can be examined in a representative reaction between an iron oxide and soluble ferrous (Fe^{+2}) iron:



Though there must be a transfer of electrons for the reaction to take place, the complete reaction is written without the electrons present:



The tendency for electron movement is referred to as the activity of the electron (Drever, 1988; Sposito, 1989). The electron activity can be measured in soil using a platinum electrode system and can be used to develop a relationship between measured potentials in soils and calculated values of ion activity.

The activity of electrons can be measured by taking the difference in potential (E) between two electrodes as in Drever's model. The potential is called Eh in

reference to the standard hydrogen electrode. The measured activity of electrons can be expressed in units of volts. The activity of the electrons is calculated using P_e , the negative log of the electron activity [-Log (e)].

P_e is related to E_h through the Nernst equation. This equation allows predictions of active redox couples at certain E_h levels:

$$P_e = F/2.303 RT E_h$$

F- Faraday's Constant= 96.42 KJ/Volt gram equiv.

R- Universal Gas Constant 8.314E-3 KJ deg⁻¹mol⁻¹

A standard potential at 25°C uses the equation:

$$E_h = E^0 + 0.059 \log (Fe^{+3}/Fe^{+2})$$

E^0 is the standard electrode potential. This is the E_h the cell would have if all the chemical species involved were in their standard states. When a calomel reference electrode is used, E^0 is calculated as $E^c + 244$ Mv. Use of a calomel electrode requires an adjustment of +244 mV to measured voltages.

The generalized equation for any reaction at 25°C is :

$$E_h = E^0 + 0.059/n \log \frac{\text{(Oxidized)}}{\text{(Reduced)}}$$

A general relationship of potentials and reduced species is given by Sposito (1989) (Table 1). Sposito names three zones in the sequence of reduction. The oxic zone is from 300-700 mV where oxygen is the dominant electron acceptor. Nitrate is

also reduced at the low end of these potentials in anaerobic microsites (Richards, 1987). The suboxic zone falls from 100-300 mV. Oxygen becomes thermodynamically unstable below 400 mV and iron and manganese become the primary electron acceptors. However, thermodynamic theory indicates that iron reduction will not occur until all the oxygen and nitrate are depleted. The anoxic conditions are those <120 mV where sulfate is the primary electron acceptor.

However, the correspondance of a measured Eh value to a specific redox couple is not absolute. There are many redox couples active at any time in the soil due to the mixed potentials that occur in a heterogeneous soil environment (Drever, 1988). Stumm and Morgan (1981) indicate that soils have localized intermediate zones of redox couples that are the result of imperfect mixing in the biological system. Bohn (1985) states that platinum electrode measurement represents the average potential of many couples and not the potential of any single couple.

These views indicate that the interpretation of measured redox potentials is very generalized and perhaps even inconclusive. However, Drever (1988) points to the fact that in waters from strongly reducing environments, minor species are often present to which the electrode responds, rendering Eh measurement of value.

Sposito (1989) concurs with Drever's observation, stating:

"In the case of flooded soil, redox reactions will be controlled by the behavior of a closed chemical system that is catalyzed effectively by bacteria and for which an equilibrium description is especially apt".

Therefore, in the highly reduced zones of flooded soils, Eh measurements as a general guide to the redox environment can be useful.

In soil morphology, manganese and iron are of primary concern because of the conspicuous colors left by their transformation. Manganese and iron are

at the more highly reduced end of Sposito's "redox ladder" (table 1), and are the most significant redox couple at potentials of 200-300 mV (Cogger and Kennedy, 1992). In this case, Eh measurement can be of considerable value in verifying the transformations of iron and manganese due to the present redoximorphic conditions in soils.

Methods

Platinum electrodes used for this project were constructed similar to those of Faulkner et al. (1989) and Szogi and Hudnall (1990). A complete description of electrode manufacture, testing and installation is available in Austin (1993) and will not be discussed here. The platinum electrodes differed slightly from those previously mentioned as they did not have a mercury junction. The electrode junction was made by soldering a copper wire lead to the platinum tip. The junction was then sealed by filling the glass tube with epoxy. This resulted in an insulated connection as well as increased strength of the electrode. Electrode data were collected weekly from October 1992 through March 1994.

Dissolved oxygen measurements were taken with an Orion model 820 self-calibrating dissolved oxygen sensor. Dissolved oxygen (DO) measurements were made directly in piezometer wells. The wells were pumped dry and allowed to refill before a reading was taken. Readings were taken by lowering the probe to the bottom of the piezometer well and then moving the probe up and down slowly until a stable reading was obtained. DO measurements were taken between December 1992 and March 1994.

However, due to sampling difficulties, only data from the second season, December 1993 to March 1994 were used for analysis.

Results and Discussion

Plot 1

Data for redox potential and duration of saturation from piezometer data are given in figure 14. Initially, the redox potentials at 25, 50 and 100 cm depths were in the 300-400 mV range. These values represented oxidizing conditions in the soil at the end of the dry season.

The potentials at the 25 cm depths dropped rapidly with the onset of saturation in November 1992. The potentials at 25 cm were the lowest throughout the period of saturation and were well below 100 mV from December 1992 to August 1993. Reduction at the 25 cm depth is intense during saturation periods. Surface saturation impedes the decomposition of organic matter by aerobic bacteria and leaves a rich source of organic carbon for anaerobic microbial reactions. The intensity of reduction is enhanced at this level by increasing soil temperature (fig. 15) coinciding with soil saturation from March to August. The increase in temperature at 25 cm corresponds to a fall in redox potential in February 1994. The increase in temperature at 25 cm is evident as the 25 cm temperature values become higher than the temperatures at the 50 and 100 cm depths to become the highest measured temperatures. The rapid rise in potentials to oxidizing conditions in September 1993 corresponded well to the end of the period of soil saturation. The potential fell

Figure 14. Duration of saturation as measured by piezometers (bottom) and electrode potentials (top) at depths of 100 cm, 50 cm, and 25 cm at Bashaw plot 1 at the Jackson-Frazier wetland in Benton County.

Figure 14.

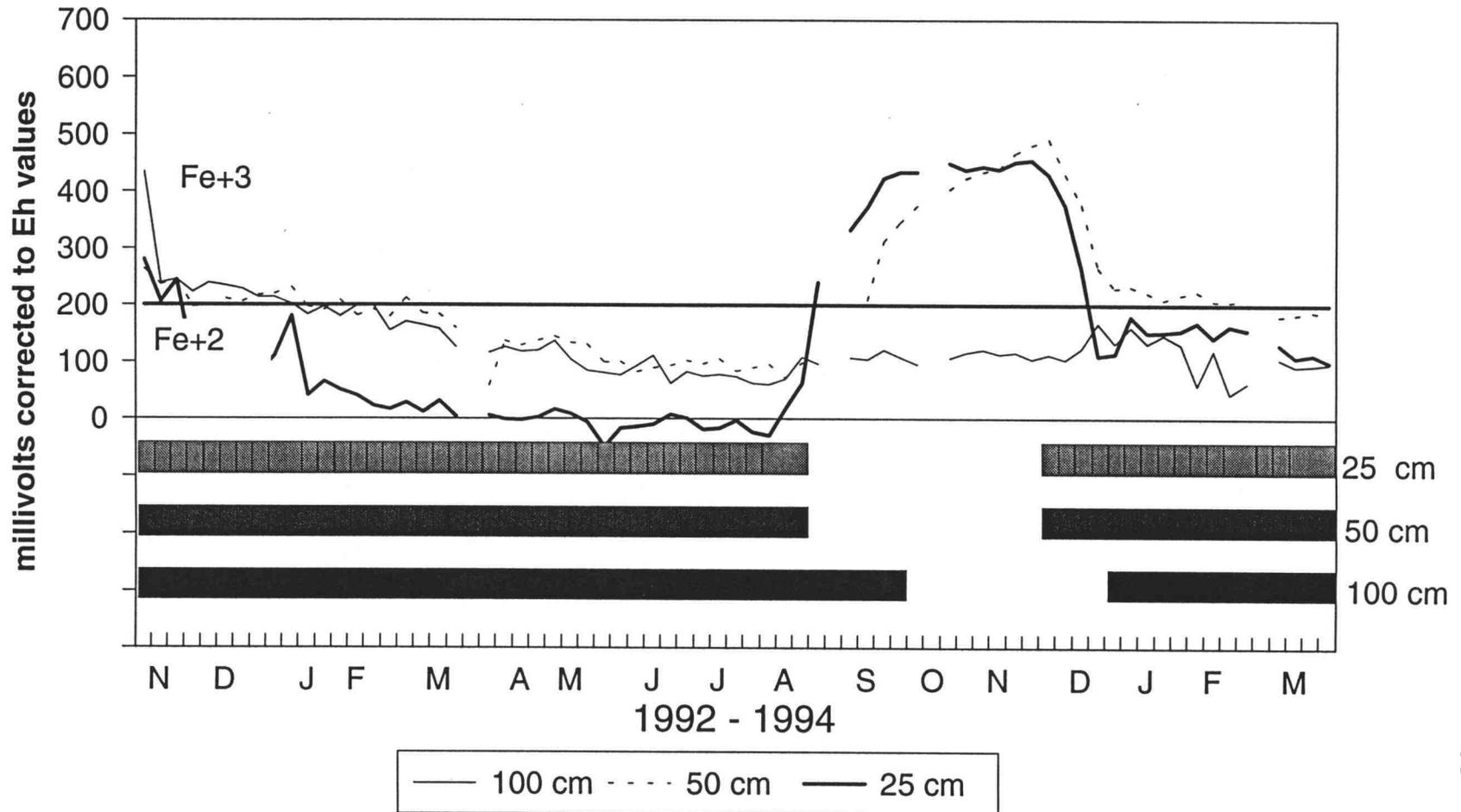
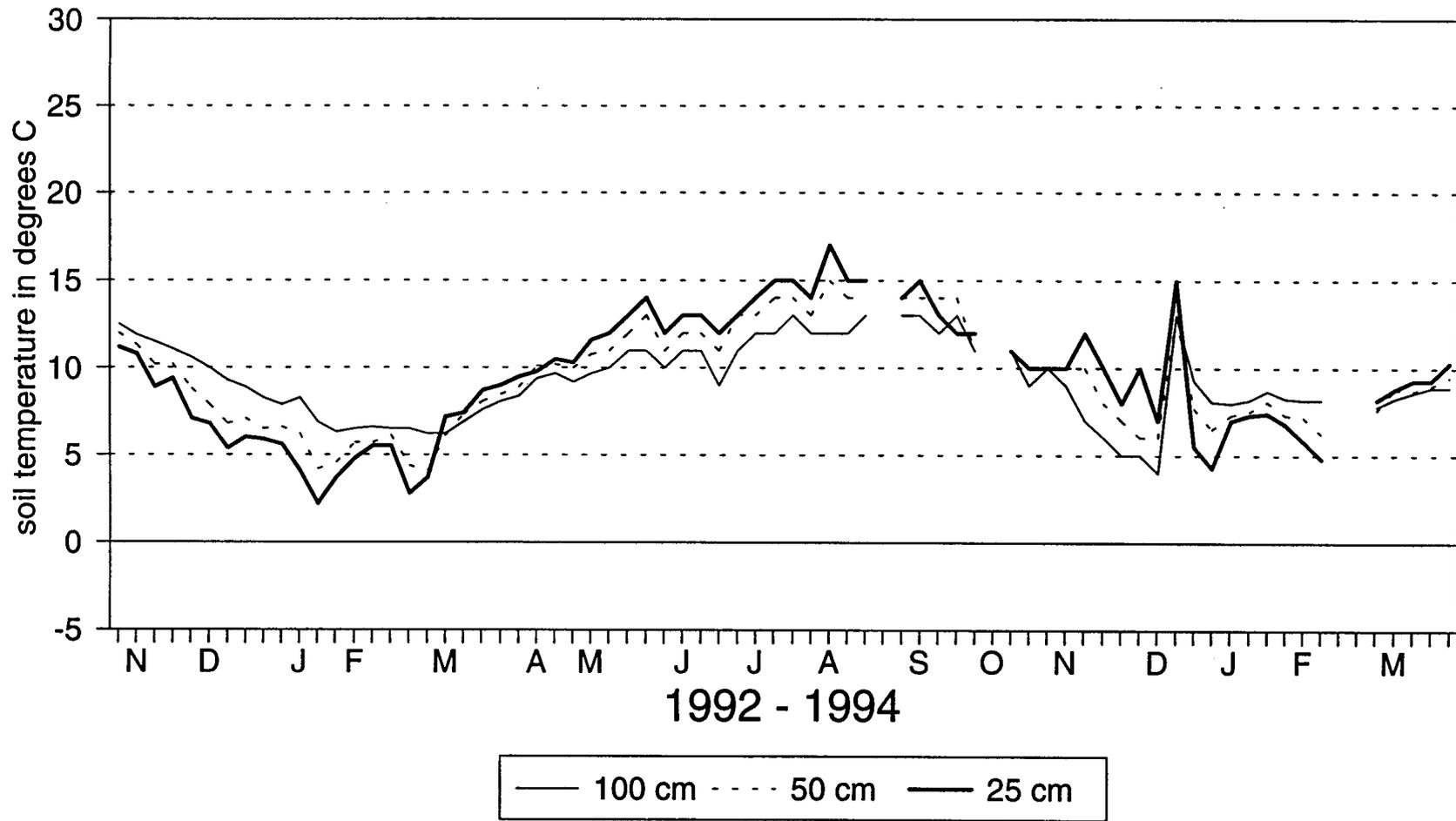


Figure 15. Soil temperature data at depths of 100 cm, 50 cm, and 25 cm at the Bashaw plot 1 at the Jackson-Frazier wetland in Benton County.

Figure 15.



again at 25 cm when the soil became saturated in December 1993 and remained there until March 1994.

The potentials at 50 and 100 cm also declined rapidly with the onset of saturation but values were not as low as for 25 cm, and they were nearly the same from November 1992 through July 1993. During this long period of saturation the potentials declined gradually from oxidizing conditions, 300-400 mV, to anaerobic conditions between 100 and 200 mV. The redox potentials at 50 and 100 cm diverged abruptly as the soil at 50 cm dried and became oxidized in August 1993, while the soil at 100 cm remained saturated.

The supply of organic matter at 50 cm because of the mixing of surface materials due to cracking during dry periods is sufficient to provide substrate for the microorganisms to drive potentials down into the anaerobic range during soil saturation. However, the redox potential responds to drying conditions in the soil which suggests that the aeration at this depth is enhanced by macropores opening as the soil dries allowing oxygen to diffuse rapidly into the soil.

In October 1993, the soil at 100 cm finally dried out, but there was no corresponding rise in potential as it remained steady at approximately 100 mV through November 1993. There was a slight rise in potential during December 1993, but renewed saturation in late December 1993 caused a noticeable drop in potential. The potential at 100 cm remained reducing until the end of the observation period. Redox potentials at 100 cm at plot 1 indicate that anaerobic processes prevailed at this depth even after the soil dried. This response indicates that aeration at this point in the soil was limited. The soil at

100 cm did not experience the same drying and cracking conditions as at 50 cm. The cracks at plot 1 may not have extended deep enough into the subsurface to allow oxygen and heat to enter the soil. The soil at plot 1 retains water late into September, and the soil remains impermeable to the diffusion of oxygen inhibiting significant aeration to that depth.

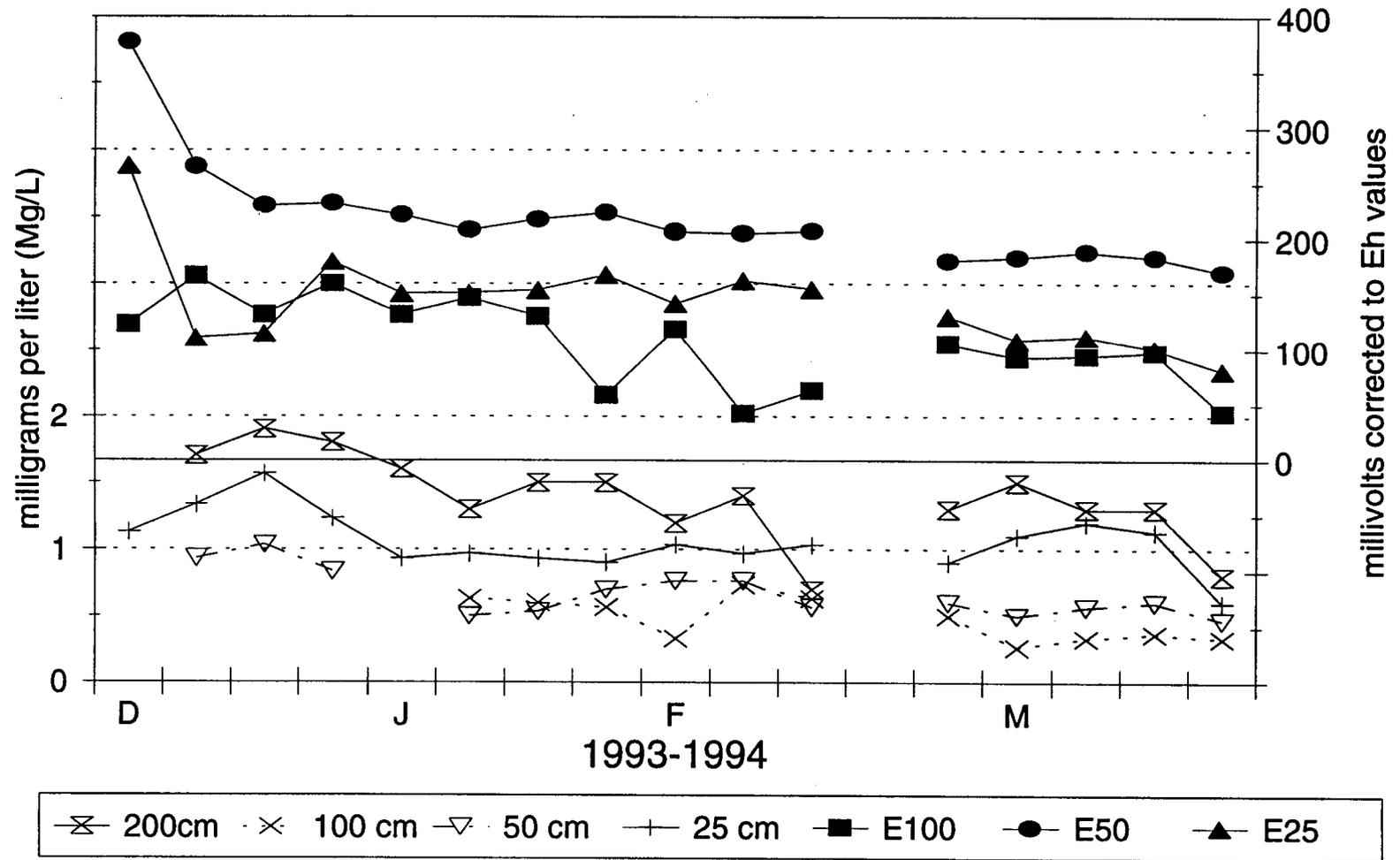
The 25 and 50 cm potentials dropped rapidly with the onset of saturated conditions in late December, 1993. The rapid drop in redox potentials indicates that the soils become saturated very quickly. Water flows down into cracks in the soil and moves outward into the soil matrix resulting in saturation soon after the first precipitation of the season. The potentials remained between 100 and 200 mV through March 1994.

Redox potentials and dissolved oxygen levels for plot 1 are displayed in figure 16. Dissolved oxygen levels at 25 cm were from 1.0 to 1.5 milligrams per liter (mg/L) in December 1993. The DO level remained at 1.0 mg/L through January and February 1994. The dissolved oxygen at 25 cm is fairly high due to the renewed supply of DO from the oxygenated water flowing over the surface. There was a slight rise in the DO level in early March 1994 followed by a sharp decline in late March 1994. The decline in DO is the result of renewed microbial activity with the increase in temperature during the early spring.

The DO levels at 50 and 100 cm were consistently below 1.0 mg/L during the monitoring period. The values were from 0.5 to 0.8 mg/L from January 1994 to March 1994. The diffusion of oxygen at these levels is impeded by the slow permeability of the soils and the blockage of many pores

Figure 16. Redox potential (top) and dissolved oxygen (bottom) as observed at the Bashaw plot 1 at the Jackson-Frazier wetland in Benton County.

Figure 16.



by water. The DO levels at 200 cm were near 1.5 mg/L throughout the monitoring period with the exception of two declines in February and March 1994. The levels here are the highest recorded and correspond to some oxygen rich source. The water flowing at 200 cm is in the silty clay loam deposit which must be connected to the hillslope baseflow at the upper edges of the wetland. Highly oxygenated water flows from the upper portions of the drainage through the subsurface deposit. There is very little demand for the oxygen by plants or microorganisms at this level so the DO is not depleted by the time it reaches the monitoring well.

Plot 2

Redox potentials and soil temperature data for plot 2 are displayed in figure 17 and figure 18 respectively. The potentials for all depths were initially in the oxidized range of 400-500 mV. The 25 and 50 cm potentials fell rapidly to 200 mV with the onset of saturation in November, 1992. Thereafter, the potentials declined slowly to 100 mV and remained at that level during the period of soil saturation from November 1992 to August 1993. The potentials at this plot are responding to the same forces of rapid saturation, reduction and increasing temperatures as at plot 1. The potentials at 25 and 50 cm also rose sharply in August 1993 in response to soil drying as cracks opened and aeration of the soil followed.

Figure 17. Duration of saturation as measured by piezometers (bottom) and electrode potentials (top) at depths of 100 cm, 50 cm, and 25 cm at Bashaw plot 2 at the Jackson-Frazier wetland in Benton County.

Figure 17.

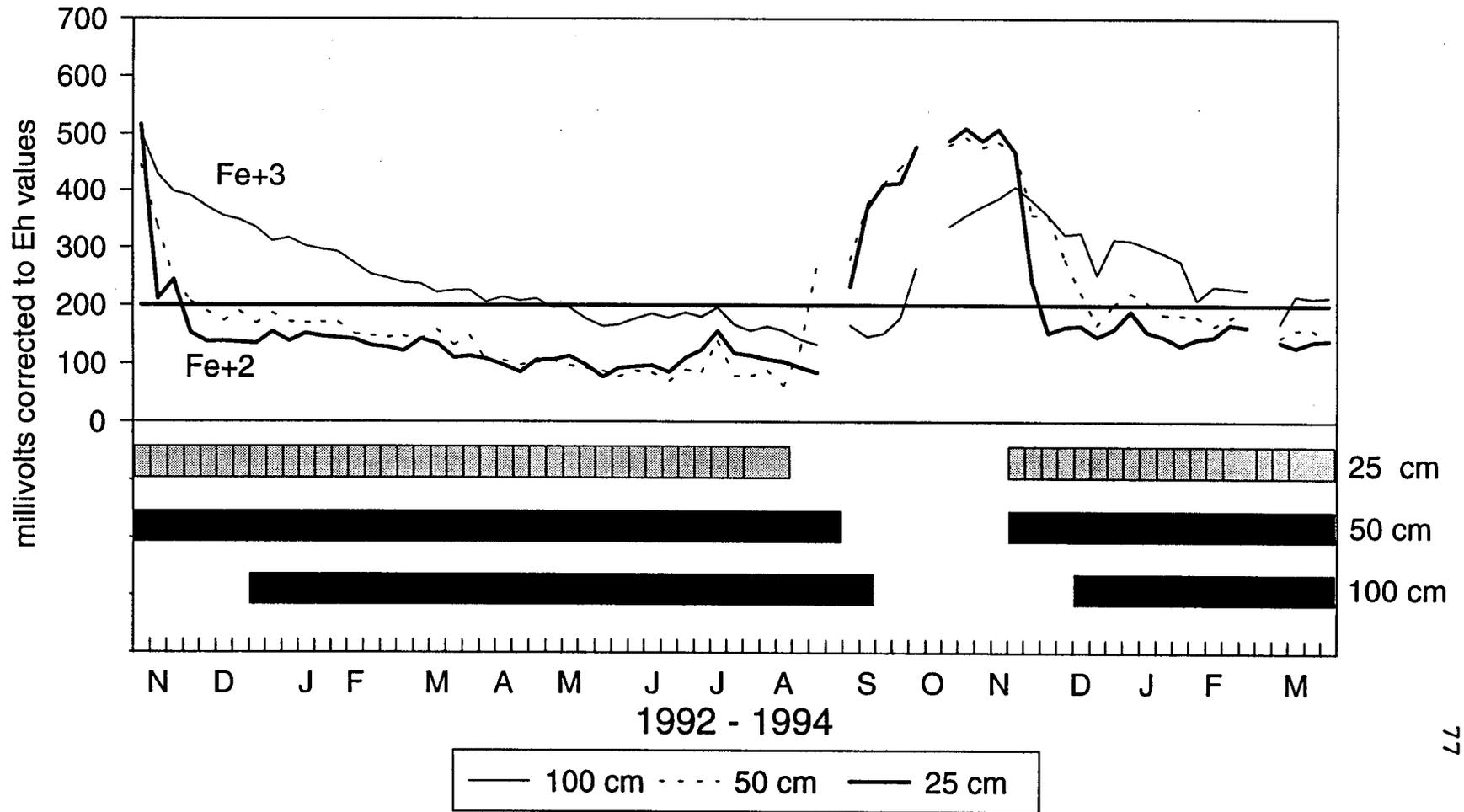
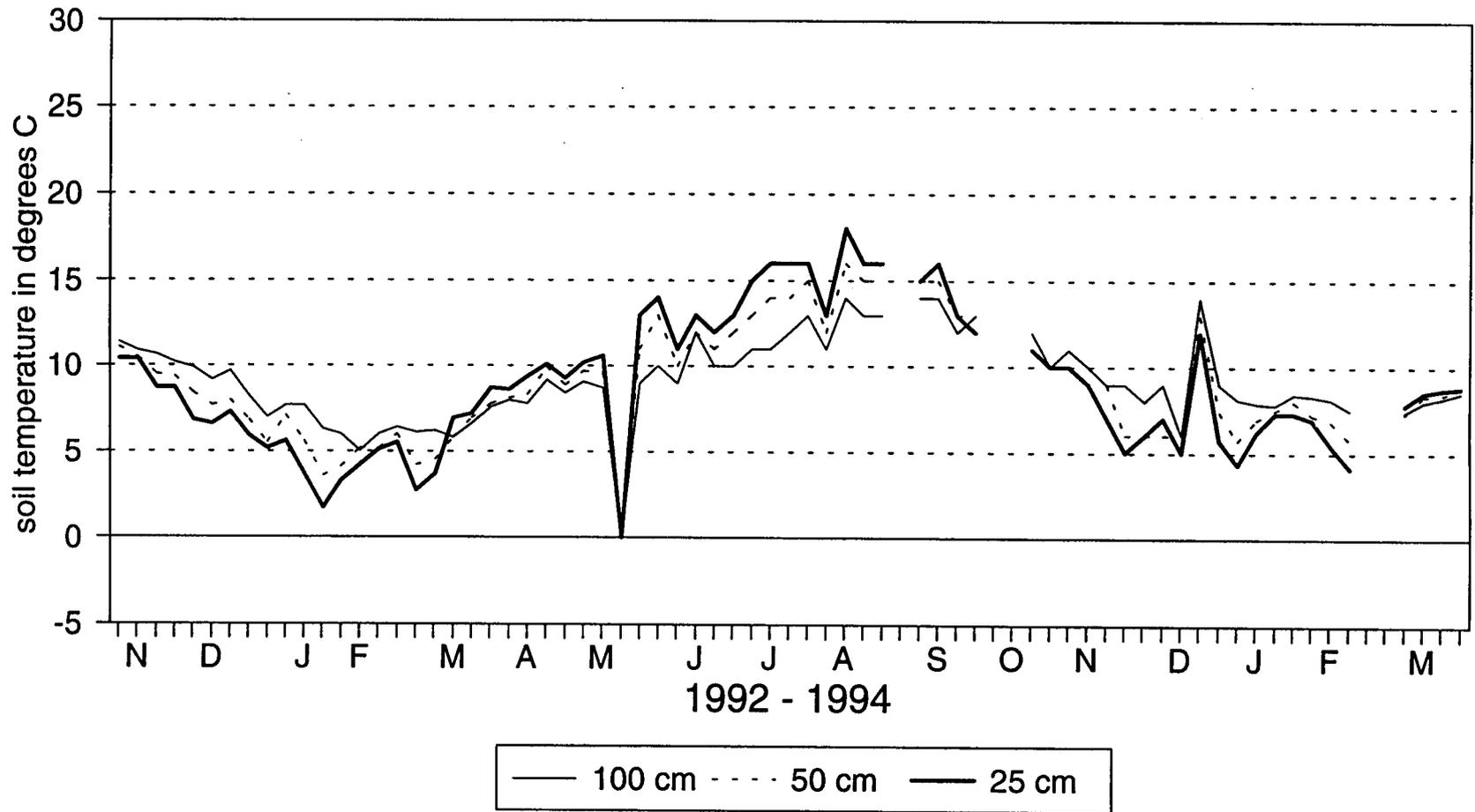


Figure 18. Soil temperature data at depths of 100 cm, 50 cm, and 25 cm at the Bashaw plot 2 at the Jackson-Frazier wetland in Benton County.

Figure 18.

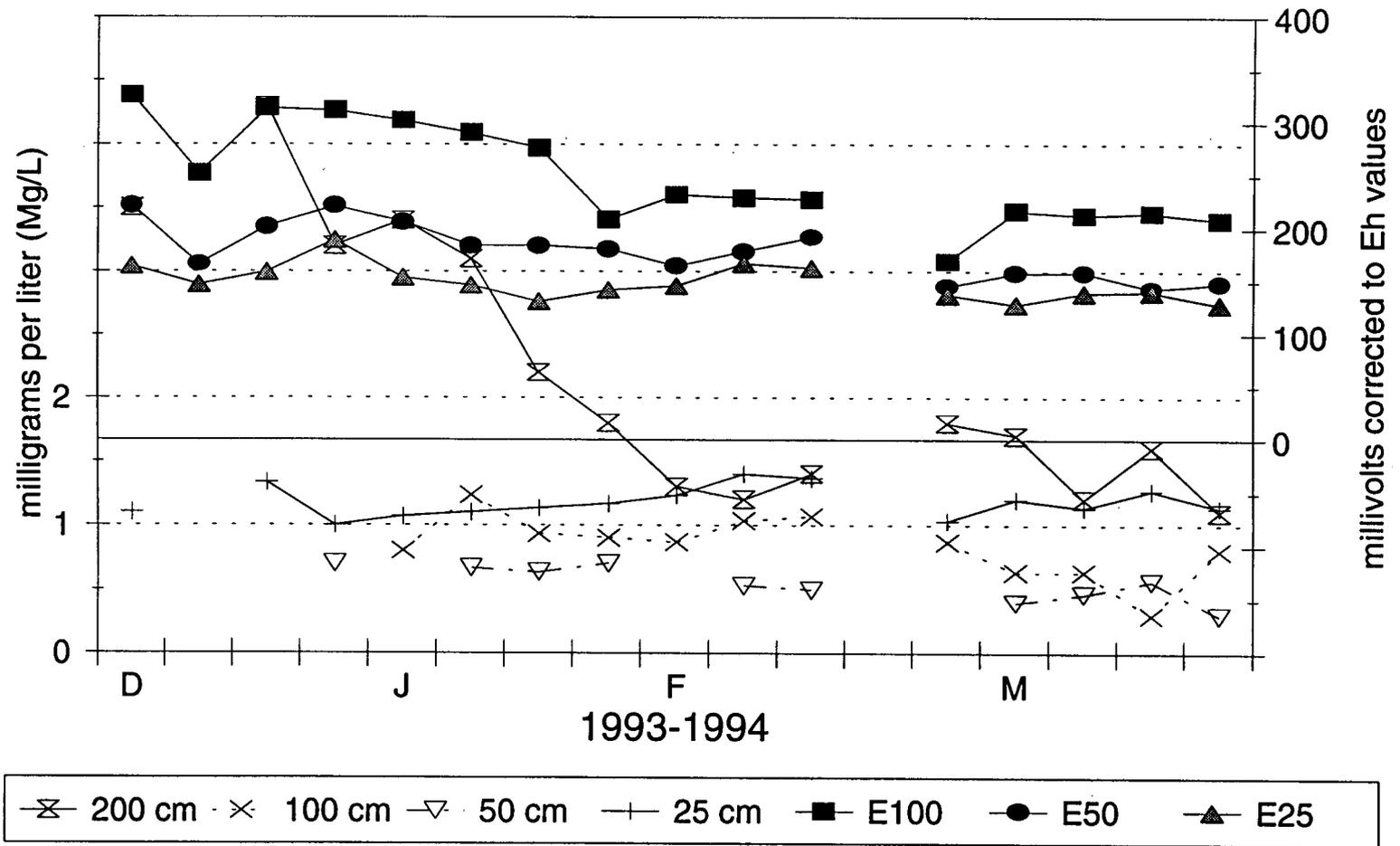


The 100 cm potentials dropped out of oxidizing conditions slowly both before and after the onset of saturation. The slow drop in redox potential corresponded to a delay in soil saturation at 100 cm. This suggests that even though the soil is not saturated, oxygen is still available at this depth, but is slowly depleted by microbial respiration and not replenished. The flow of oxygen is impeded due to both the water above which is ponded on the surface, and the water moving through the silty clay loam deposit below the 100 cm electrodes. The soil was saturated at 100 cm in December 1992. Potential at 100 cm rose in September 1993 in response to oxidizing conditions in the soil.

The 100 cm potential at plot 2 is well correlated with the end of saturation, as the potential rises rapidly once the soil has no free water. The 100 cm potential at plot 2 also remained above 200 mV for a longer period of time than the 100 cm potentials at plots 1 and 3. The soil at this plot may not have as much organic matter present, inhibiting microbial reduction. More likely, however, is the influence of the silts underlying the black clay layer. The DO level for 100 cm at plot 2 is higher than at plot 1 (figs. 16,19). The 200 cm DO level at plot 2 ranged from high of 4 mg/L in late December 1993 to 1.2 mg/L in February 1994. The level fluctuated between 1.2 and 1.8 mg/L between late February to March 1994. The 100 cm DO level ranged from 0.9 to 1.2 mg/L in January and February 1994 and then fell to 0.9 to 0.5 mg/L in late February and March 1994 as microbial action depleted the supply of oxygen with rising temperatures.

Figure 19. Redox potential (top) and dissolved oxygen (bottom) as observed at the Bashaw plot 2 at the Jackson-Frazier wetland in Benton County.

Figure 19.



The higher DO levels at 100 cm may be due to the shallower contact with the oxygen rich silts at plot 2. The oxygen may be diffusing upward from the silty clay loam aquifer into the aquitard. The DO measurements cannot be taken as absolutely corresponding to the redox environment, but as a relative measure they may indicate that more oxygen is influencing the redox potentials at plot 2.

The potentials at 25 and 50 cm again fell rapidly with the onset of soil saturation in November 1993, and then remained level at about 150 mV through March 1994. The 100 cm potentials showed the same response as in 1992-93 with a slow decline following the later soil saturation at 100 cm due to the trapping of oxygen and its slow decline due to microbial activity. The 100 cm potentials were measured at 200 mV in March 1994.

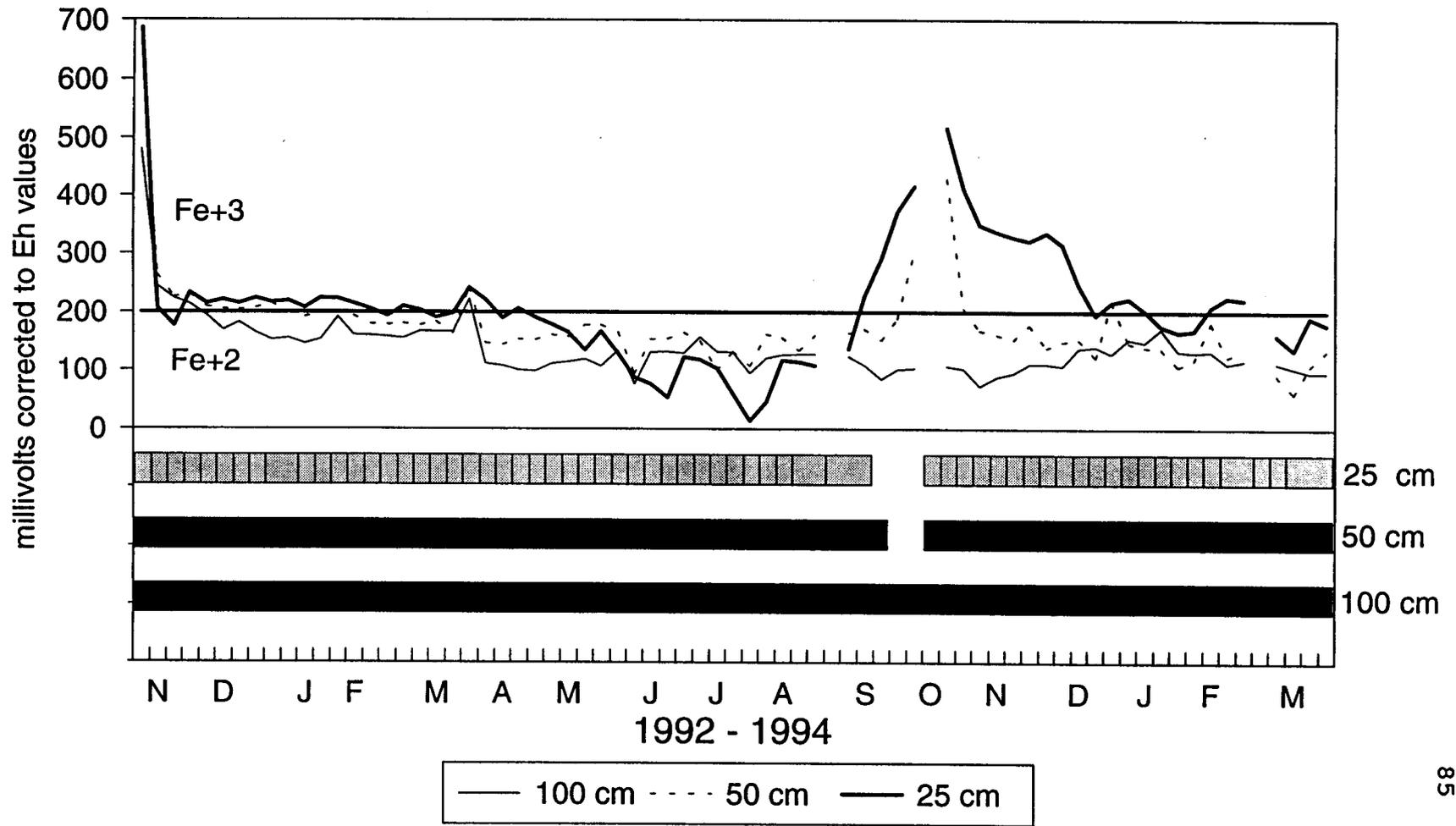
The DO level at 25 cm was steady between 1.0 and 1.5 mg/L throughout the monitoring period corresponding to oxygenated surface water recharging this depth with DO. The DO level at 50 cm was between 0.5 and 0.8 mg/L over the same time span indicating impedance to the diffusion of oxygen by the black clay.

Plot 3

The redox potentials at all depths at plot 3 (fig. 20) were in the oxidizing range at the first observation in late October 1992. All potentials fell rapidly after initial soil saturation and remained relatively constant at about 200 mV until March 1993, after which they declined to levels below 200 mV. The 25 and 50 cm potentials dropped to 100 mV in May 1993, and the 25 cm potential alternately fell to 0 mV and rose to 100 mV from June through August 1993.

Figure 20. Duration of saturation as measured by piezometers (bottom) and electrode potentials (top) at depths of 100 cm, 50 cm, and 25 cm at the Cove plot 3 at the Jackson-Frazier wetland in Benton County.

Figure 20.



The potentials at 25 and 50 cm rose to oxidizing conditions (300–400 mV) in response to soil drying in September 1993. The potentials remained in the oxidizing range in September and October 1993 after which they started to fall into the anaerobic range again. The 50 cm potential dropped rapidly to 150 mV in October 1993 and remained steady between 100 and 200 mV from November 1993 to March 1994. The 25 cm potential dropped from oxidizing potentials in October 1993 to reducing potentials (200 mV) in December 1993. The potentials at 25 cm fluctuated between 100 mV and 225 mV from December 1993 to March 1994.

The 100 cm potentials never rose to oxidizing conditions throughout the period of observation. Anaerobic potentials corresponded to the indication that the soil remained saturated throughout the observation period. The potentials show an upward trend from October to January 1994. This upward movement may be due to the soil temperature at this depth decreasing (fig. 21) leading to less microbial activity and the generation of less electrons for the reduction of manganese and iron. Electrode potentials at 100 cm at plot 3 are well correlated with the permanent saturation condition. The long term saturation may be the result of the slow downward movement of water from the 25 and 50 cm levels to the 100 cm level. A rise in redox potential in November indicates that the reduction is moderated by a drop in soil temperature which slows microbial activity.

Redox potentials and dissolved oxygen levels for plot 3 are displayed in figure 22. Dissolved oxygen measurements were only collected at 25 to 50 cm because the slow recharge at 100 and 200 cm made obtaining accurate

Figure 21. Soil temperature data at depths of 100 cm, 50 cm, and 25 cm at the Cove plot 3 at the Jackson-Frazier wetland in Benton County.

Figure 21.

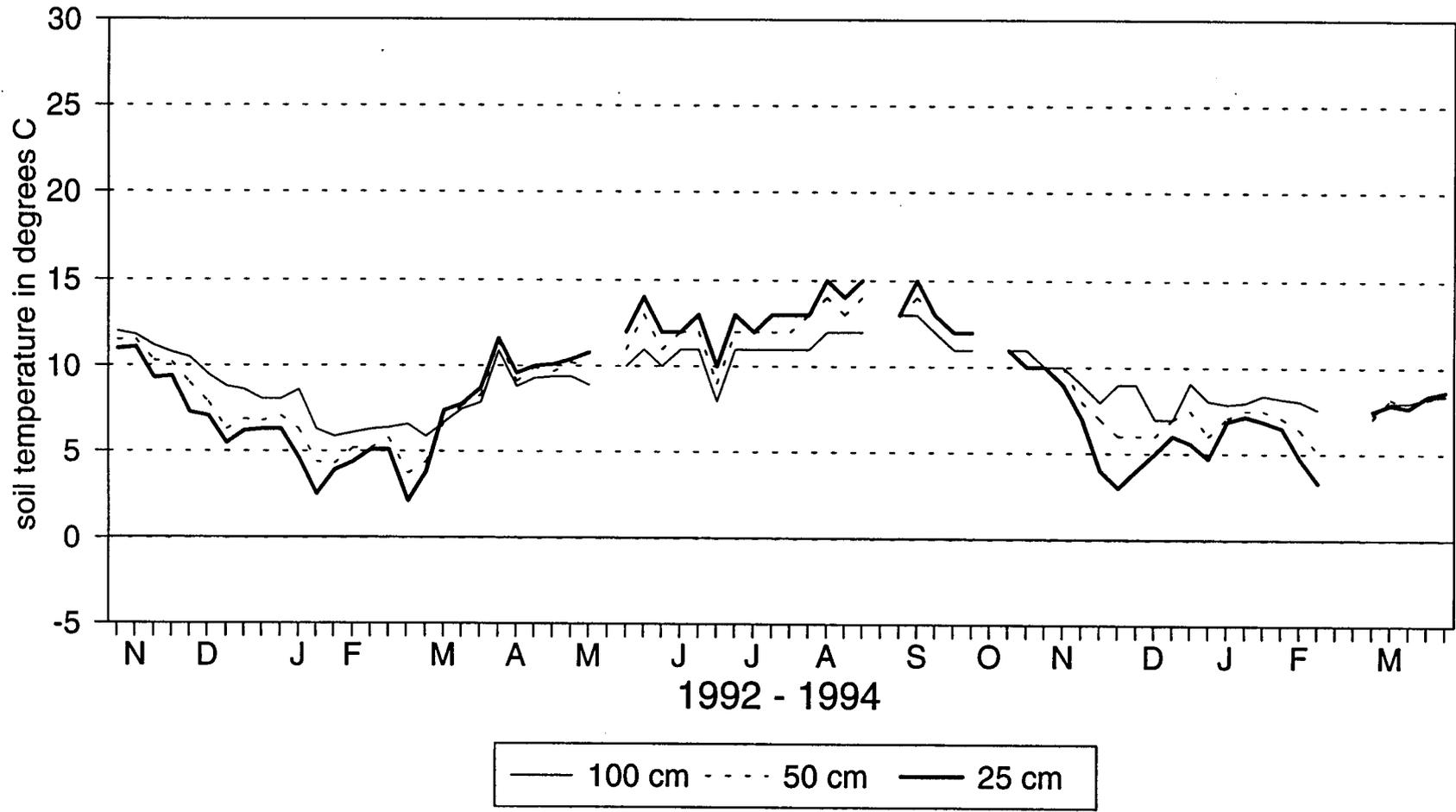
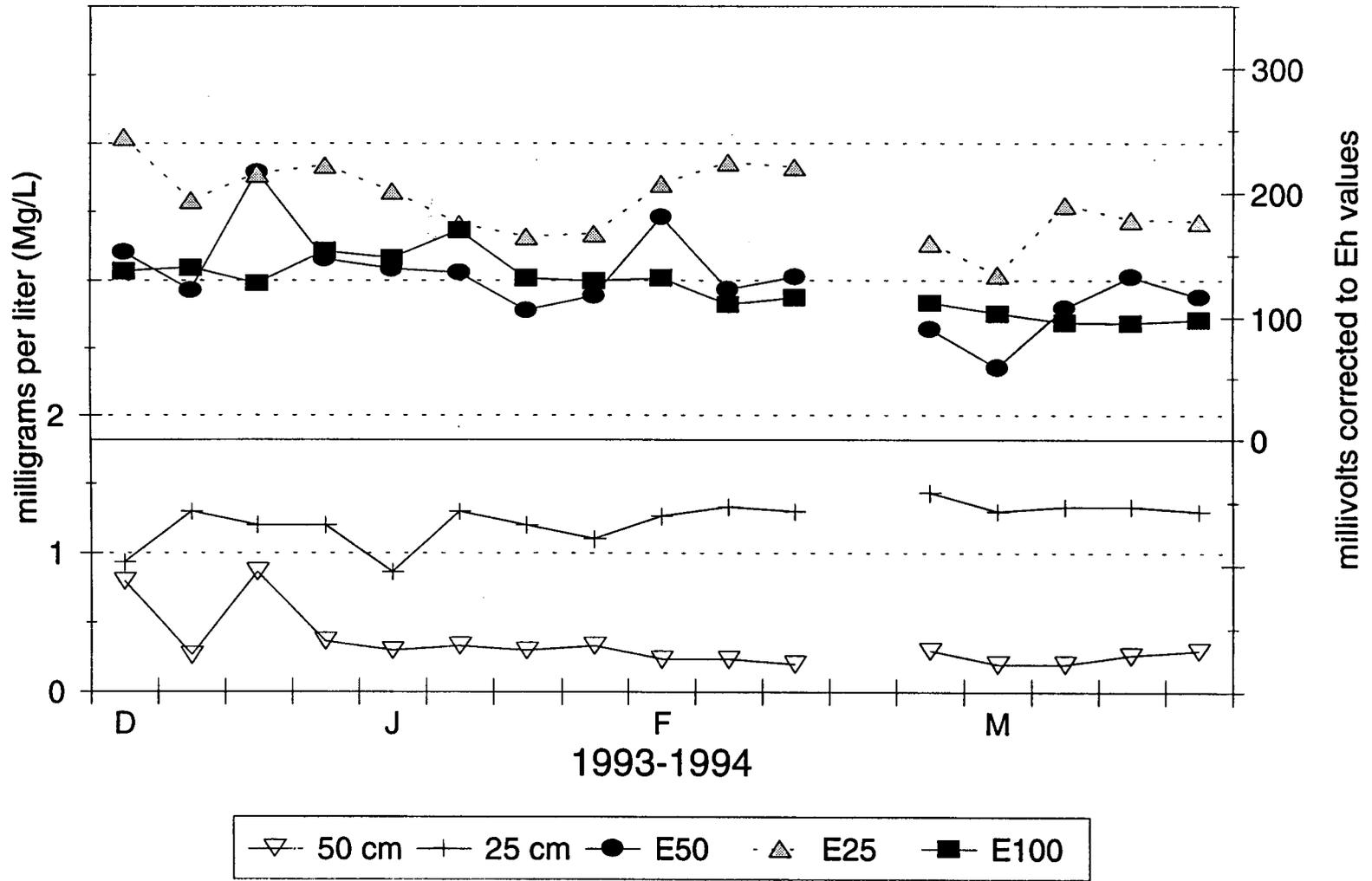


Figure 22. Redox potential (top) and dissolved oxygen (bottom) as observed at the Cove plot 3 at the Jackson-Frazier wetland in Benton County.

Figure 22.



readings very difficult. The DO level at 25 cm was between 0.9 and 1.4 mg/L for the duration of observation from December 1993 to March 1994. The DO levels did not fluctuate much during this period. The 50 cm DO measurements alternated in December 1993 from 0.8 to 0.4 mg/L but became steady at 0.4 mg/L from January to March 1994. These data are consistent with the observations from plots 1 and 2 with the 25 cm values remaining higher because of abundant oxygen in surface water and the 50 cm DO values remaining low because of the impeded diffusion of oxygen.

Abundant organic matter and the extended period of saturation lead to the drop of potentials into the iron reduction range of 100-200 mV at all depths at all plots. These conditions suggest that the potentials would continue to drop and move into the reduction of sulfate and even methane. Except for the 25 cm potentials at plot 1, however, this did not happen. This may be attributed to an abundant supply of Fe^{+3} which poises the redox system at 100 mV.

Reducible iron is available in the sediments that are transported from the surrounding landscape. Another possibility for the poisoning of the redox system may be the diffusion of oxygen upward from the relatively oxygen rich waters moving through the silt strata at 2 meters in depth. The oxygen may influence several of the electrodes, leading to higher average potentials.

The long term reduction of the soil may result in the mobilization and loss of iron which is reflected as low chroma colors in the matrix. The colors for the Bss horizons show some reduced colors (10YR 4/1, 2.5Y 4/2) indicating that there may be some iron removal taking place in certain portions of the soil matrix. However, due to the shrink swell nature of the soil, dark surface

material falls into cracks when the soil is dry and mixes with the subsurface soil horizons, making accurate interpretations of soil color patterns difficult.

Reduced colors in the soil matrix are visible, but may be less distinct due to the mixing with the dark surface material.

Concentrations of iron and manganese in the soils are manifest as abundant rounded iron-manganese concretions throughout the Bss horizons. These concretions are a sink for the precipitation of soluble ferrous iron (Fe^{+2}) when the soil becomes oxidized in the late summer to fall. However, these concretions may also be relict features which have been transported into the soils from the surrounding foothills. Schwertmann and Fanning (1976) conclude that optimal concretion formation occurs in soils with rapid changes in aeration and fluctuations in redox potential. The data for the Jackson-Frazier soils indicate that the reduction in the soil is long term with little fluctuation between oxidizing and reducing conditions. However, Jackson-Frazier had been previously drained earlier this century and the conditions may have been adequate for concretion formation. The prevalence of beaver in the wetland maintains a ponded environment which may be different than the recent decades of development at Jackson-Frazier.

Mobile reduced iron may also move downward or even laterally through the reduced soils and precipitate in areas of higher oxygen concentration. Brown (7.5yr 4/4) iron masses lower in the horizons at plots 1 and 2 provide evidence that the iron solubilized in the upper Bss horizons may have moved downward with water or diffused along a concentration gradient and precipitated at the contact with the underlying silty material. Dissolved oxygen

measurements showed that there were relatively higher levels of dissolved oxygen at the contact with the 2C horizon (Irish Bend Silt) than in the Bss horizons.

The textural break between the overlying black clay horizon and the silty clay and silty clay loam creates a very unique situation for redox conditions. Iron can accumulate as high chroma masses in fine textured horizons overlying coarse textured horizons under the appropriate reducing conditions (Clothier et al., 1978; Veneman et al., 1976). Maximum concentrations of iron have been found in somewhat poorly drained soils, with fewer concretions in very poorly drained conditions in similar landscapes (Richardson and Hole, 1979). The silts have relatively higher fluxes of dissolved oxygen flowing through them than the overlying clay horizons, which may induce a redox potential change and cause iron to become concentrated in this zone. If the oxygen concentration at depth is high enough, the oxygen may diffuse upward into the clay horizons along a concentration gradient to the lower concentrations of oxygen in the clay horizons of the soil. The oxygen may cause a localized change in the redox potential leading to the concentration of precipitated iron as concretions in the zone of smaller pores. This may explain why concretions are forming during extended periods of saturation.

Another possibility for the presence of oxygen in the clay horizons is the fact that water shuts off the soil from the surface and sub-surface creating a zone of trapped oxygen for some time. The oxygen may not diffuse out of the soil, but may be sequestered in small pores creating micro-environments for the precipitation of iron and manganese.

Literature Cited

- Allison, I.A. 1953. Geology of the Albany Quadrangle, Oregon. State of Oregon. Oregon Dept. of Geol. and Min. Ind. Bull. 37.
- Austin, W. 1994. Duration of saturation and redox potentials in selected Willamette Valley soils. M.S. thesis, Oregon State University, Corvallis, OR. 256 pp.
- Baldwin, E.M. 1964. Geology of the Dallas and Valsetz quadrangles, Oregon. Oregon Dept. of Geol. and Min Ind. Bull. 35(revised).
- Balster, C.A. and R.B. Parsons. 1968. Geomorphology and Soils, Willamette Valley, Or. Or. Agric. Exp. Stn. Spec. Rep. 265.
- Balster, C.A. and R.B. Parsons. 1969. Late Pleistocene stratigraphy, southern Willamette Valley, Oregon. Northwest Sci. 43:116-129.
- Bartlett, R. 1986. Soil redox behavior. Pp. 179-208 *In* Soil Physical Chemistry. Sparks, D.L. (ed.) CRC Press, Boca Raton, FL. 308 pp.
- Bohn, H., B. McNeal, and G. O'Connor. 1985. Soil Chemistry. 2nd. ed. Wiley and Sons, New York. 276 pp.
- Bouma, J. 1990. Report of ICOMAQ. Circular 10. *In*: Eighth International Soil Correlation Meeting, Classification and Management of Wet Soils. U.S.D.A., Soil Conservation Service, National Soil Survey Center, Lincoln, NE.
- Brinkman, R. 1979. Ferrollysis, a soil-forming process in hydromorphic conditions. Agricultural Res. Rep. 887, Center for Agricultural Publishing and Documentation, Wageningen.
- Buffkin-Drost, M. 1985. Preliminary Investigation into the hydrology of Jackson-Frazier wetland. Report at the Dept. of Geography, Oregon State University, Corvallis, OR. 42 pp.
- Clothier, B.E., J.A. Pollok, and D.R. Scotter. 1978. Mottling in soil profiles containing a coarse-textured horizon. Soil Sci. Soc. Am. J. 42:761-763.
- Cogger, C.G. and P.E. Kennedy. 1992. Seasonally saturated soils in the Puget lowland I. Saturation, reduction, and color patterns. Soil Science 153:421-433.
- Couto, W., C. Sanzonowicz, and A. de O. Barcellos. 1985. Factors affecting oxidation-reduction processes in an Oxisol with a seasonal water table. Soil Sci. Soc. Am. J. 49:1245-1248.

- 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, Wash. D.C. FWS/OBS-79/31.
- Daniels, R.B. and R.D. Hammer. 1992. *Soil Geomorphology*. J. Wiley and Sons, Inc., New York. 236 pp.
- Daniels, R.B., E.E. Gamble, and S.W. Buol. 1973. Oxygen content in the ground water of some North Carolina Aquults and Udults. pp. 153-166. *In* R.R. Bruce, K.W. Flach, and H.M. Taylor (eds.). *Field Soil Water Regime*. SSSA Spec. Publ. No. 5, Soil Sci. Soc. Am., Madison, WI.
- Domenico, P.A. and F.W. Schwartz. 1990. *Physical and Chemical Hydrogeology*. John Wiley and Sons, Inc. New York. 824 pp.
- Drever, J.I. 1988. *The Geochemistry of Natural Waters*. Prentice Hall, Inc. Englewood Cliffs, N.J. 437 pp.
- Dudal, R. 1992. Introductory Address: Wet Soils pp. 1-4 *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.
- Environmental Laboratory. 1987. *Corps of Engineers Wetland Delineation Manual*. U.S. Army Engineer Waterways Experiment Station, Vicksburg, MS. Tech. Rpt. Y-87-1.
- Fanning, D.S. and R.B. Fanning. 1989. *Soil: Morphology, Genesis, and Classification*. Wiley and Sons, Inc. New York.
- Faulkner, S.P., W.H. Patrick, and R.P. Gambrell. 1989. Field techniques for measuring wetland soil parameters. *Soil Sci. Soc. of Am. J.* 53:883-890.
- Federal Interagency Committee for Wetland 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 U.S.D.A. Soil Conservation Service, Washington, D.C.
- Franzmeier, D.P., J.E. Yahner, G.C. Steinhardt, and H.R. Sinclair, Jr. 1983. Color patterns and water table levels in some Indiana soils. *Soil Sci. Sci. Am. J.* 47:1196-1202.
- Freeze, R.A. and J.A. Cherry. 1979. *Groundwater*. Prentice Hall Inc., Englewood Cliffs, New Jersey. 604 pp.
- Gelderman, F.W. 1970. *Willamette and Associated Soils on Late Pleistocene Geomorphic Surfaces Polk County, Oregon*. M.S. Thesis, Oregon State University, Corvallis, OR. 181 pp.

- Gerrard, J. 1992. *Soil Geomorphology: An Integration of Pedology and Geomorphology*. Chapman and Hall, London. 269 pp.
- Glasmann, J.R. and G.F. Kling. 1980. Origin of soil materials in foothills of the Willamette Valley, Oregon. *Soil Sci. Soc. Am. J.* 44:123-130.
- Glasmann, J.R., R.B. Brown, and G.F. Kling. 1980. Soil-Geomorphic relationships in the western margin of the Willamette Valley, Oregon. *Soil Sci. Soc. Am. J.* 44:1045-1052.
- Marshall, J. 1985. *A Value Assessment of the Jackson-Frazier wetland*. MS thesis, Oregon State University, Corvallis, OR.
- Mitsch, W.J. and J.G. Gosselink. 1986. *Wetlands*. Van Nostrand Reinhold. New York.
- Pickering, E.W. and P.L.M. Veneman. 1984. Moisture regimes and morphological characteristics in a hydrosequence in central Massachusetts. *Soil Sci. Soc. Am. J.* 48:113-118.
- Richards, B.N. 1987. *The Microbiology of Terrestrial Ecosystems*. John Wiley and Sons, Inc. New York. 399 pp.
- 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.
- Rowell, D.L. 1981. Oxidation and Reduction. Chapter 7 *In* Greenland, D.J. and M.H.B. Hayes, eds., *The Chemistry of Soil Processes*. John Wiley and Sons, Ltd.
- Schwertmann, U. and D.S. Fanning. 1976. Iron and manganese concretions in hydrosequences of soils in loess in Bavaria. *Soil Sci. Soc. Am. J.* 40:731-738.
- Schwertmann, U. and R.M. Taylor. 1989. Iron Oxides. pp. 379-437 *In*: J.B. Dixon and S.B. Weed (eds.) *Minerals in Soil Environments*. Soil Sci. Soc. Am., Madison.
- Soil Survey Division Staff. 1993. *Soil Survey Manual*. Agricultural Handbook 436. Government Printing office, Washington, DC.
- Soil Survey Staff. 1992. *Keys to Soil Taxonomy*. 5th edition. SMSS Technical Monograph No. 19. Blacksburg, VA: Pocohontas Press, Inc. 556 pp.
- Sposito, G. 1989. *The Chemistry of Soils*. Oxford University Press, New York.
- Stumm, W. and J.J. Morgan. 1981. *Aquatic Chemistry*, 2nd ed. Wiley-Interscience, New York. 780 pp.

- Szogi, A.A. and W.H. Hudnall. 1990. Measurement of redox potentials in soils with permanently installed platinum electrodes. Manuscript 91-09-5227. Louisiana Agricul. Exp. Stn.
- Theissen, A.A. and M.E. Harward. 1962. A paste method for preparation of slides for clay mineral identification by X-ray diffraction. *Soil Sci. Soc. Am. Proc.* 26:90-91.
- Tiner, R. W. 1991. The concept of a hydrophyte for wetland identification. *BioScience* 41:236-247.
- 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.
- Veneman, P.L.M. and E.W. Pickering. 1983. Salt bridge for field redox potential measurements. *Commun. in Soil Sci. Plant Anal.* 14(8):669-677.
- Vepraskas, M.J. and L.P. Wilding. 1983. Aquic moisture regimes in soils with and without low chroma colors. *Soil Sci. Soc. Am. J.* 47:280-285.
- Vepraskas, M.J. 1992. Redoximorphic features for identifying aquic conditions. Technical Bulletin 301. North Carolina Agricultural Research Service, North Carolina State University, Raleigh, North Carolina.
- Vepraskas, M.J. and W.R. Guertal. 1992. Morphological indicators of soil wetness. Pp. 307-312 *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.

APPENDICES

APPENDIX A: SOIL PROFILE DESCRIPTIONS

Jackson-Frazier
Plot I

- A1 0 to 2 cm; 80 percent very dark grayish brown (10YR 3/2) mucky silt loam, light gray (10YR 7/2) dry; few fine prominent strong brown (7.5YR 5/6) mottles; weak very fine granular structure; soft, very friable, nonsticky, nonplastic; many very fine and fine roots; many very fine interstitial pores; clear smooth boundary. The upper 2 cm is probably a deposit of fine volcanic ash.
- A2 2 to 8 cm; 50 percent very dark grayish brown (10YR 3/2) mucky silt loam, grayish brown (10YR 5/2) dry; many fine prominent strong brown (7.5YR 5/6) mottles; weak very fine granular structure; soft, very friable, nonsticky, nonplastic; many very fine and fine roots; many very fine interstitial pores; clear smooth boundary.
- A3 8-15 cm; 50% very dark gray (10YR 3/1) silty clay loam, dark gray (10YR 4/1) dry; many fine and medium prominent strong brown (7.5YR 5/6) mottles; moderate fine subangular blocky structure; hard, friable, sticky, plastic; few very fine and fine roots; common very fine intersitial pores; few faint patchy black (10YR 2/1) manganese or iron-manganese stains; abrupt smooth boundary.
- A4 15 to 41 cm; 95 percent black (10YR 2/1) clay, dark gray (10YR 4/1) dry; few fine distinct brown (7.5YR 5/4) mottles; moderate coarse prismatic parting to strong medium subangular blocky structure; extremely hard, extremely firm, very sticky, very plastic; few very fine and fine roots; few very fine tubular pores; few fine rounded iron-manganese concretions; gradual wavy boundary. Vertical cracks 3 to 4 inches apart.
- Bss1 41 to 66 cm; 60 percent black (10YR 2/1) and 35 percent dark grayish brown (10YR 4/2) clay, dark gray (10YR 4/1) and light brownish gray (10YR 6/2) dry; few fine distinct brown (7.5 YR 5/4) mottles; moderate coarse prismatic parting to strong medium subangular blocky structure; extremely hard, extremely firm, very sticky, very plastic; few very fine roots; few very fine tubular pores; few distinct continuous intersecting slickensides; common fine rounded iron-manganese concretions; gradual wavy boundary.
- Bss2 66 to 104 cm; 60 percent dark grayish brown (10YR 4/2) clay, light brownish gray (10YR 6/2) dry; moderate coarse prismatic parting to strong medium angular blocky structure; extremely hard, extremely firm, very sticky, very plastic; few very fine tubular pores; many prominent continuous black (10YR 2/1) organic coats on faces of pedes and in pores and many prominent continuous intersecting slickensides; gradual wavy boundary.
- Bss3 104 to 119 cm; 80 percent dark grayish brown (2.5 Y 4/2) clay, light brownish gray (2.5Y 6/2) dry; few fine distinct brown (7.5YR 4/4) mottles;

moderate coarse prismatic parting to strong medium angular blocky structure; extremely hard, extremely firm, very sticky, very plastic; few very fine tubular pores; many prominent continuous black (10YR 2/1) organic coats on faces of peds and in pores and common prominent continuous intersecting slickensides; clear wavy boundary.

- 2C1 119 to 135 cm; grayish brown (2.5Y 5/2) and dark grayish brown (2.5Y 4/2) silty clay, light gray (2.5Y 7/2) dry; few fine distinct brown (7.5YR 4/4) mottles; massive; extremely hard, extremely firm, very sticky, very plastic; few fine tubular pores; few prominent continuous black (10YR 2/1) organic coats in root channels and/or pores; gradual wavy boundary.
- 2C2 135 to 155 cm; olive brown (2.5Y 4/4) silty clay loam, light gray (2.5Y 7/2) dry; few fine distinct brown (7.5YR 4/4) mottles; massive; very hard, firm, very sticky, plastic; few fine tubular pores; few prominent continuous very dark gray (10YR 3/1) organic coats in root channels and/or pores.
- 2C3 155 to 251 cm; silty clay loam; massive; hard, firm, slightly sticky, slightly plastic. Material collected with auger.
- 2C4 251 to 290 cm; silty clay loam; massive; hard, firm, slightly sticky, slightly plastic. Material collected with auger.

Jackson Frazier
Plot II

- A1 0-18 cm. Very dark brown (10YR 2/2) silt loam; common fine prominent yellowish red (5YR 4/6) mottles; strong medium granular structure; friable, slightly sticky and plastic; many very fine, many fine and common medium roots; few medium concretions; clear smooth boundary.
- Bss1 18-34 cm. Very dark gray (N3/0) silty clay; few fine faint reddish brown (5YR 4/4) mottles; strong, very coarse prismatic structure; firm, very sticky and plastic; few fine and common very fine roots; few medium concretions; common distinct slickensides; gradual smooth boundary.
- Bss2 34-50 cm. Black (N2/0) clay; few fine faint reddish brown (5YR 4/4) mottles; strong coarse to very coarse prismatic structure; vertical cracks 10 mm wide and pressure faces on vertical surfaces are distinct; firm, very sticky and plastic; few fine and few very fine roots; common medium black concretions; common distinct slickensides; gradual smooth boundary.
- Bss3 50-82 cm. Dark grayish brown (2.5Y 4/2) and very dark gray (N3/0) silty clay; moderate medium to coarse prismatic structure; vertical cracks 10 mm wide and pressure faces on vertical surfaces are distinct; firm, sticky and plastic; few fine and few very fine roots; common medium and coarse black concretions; many prominent slickensides; gradual smooth boundary.
- 2Btb 82-100 cm. Light olive brown (2.5Y 5/4) and very dark gray (N3/0) silty clay; few fine faint strong brown (7.5YR 4/6) mottles; medium subangular blocky structure; firm, sticky and plastic; few fine roots; many common distinct clay films on vertical ped faces; many coarse black concretions; clear smooth boundary.
- 2BC1 100-118 cm. Light olive brown (2.5Y 5/4) silty clay; common fine distinct light brown (7.5YR 6/4) mottles; weak medium subangular blocky structure; firm, slightly sticky and slightly plastic; few fine roots; few medium black concretions; few distinct clay films lining large pores; clear smooth boundary.
- 2BC2 118-140 cm. Dark grayish brown (2.5Y 4/2) silty clay; common fine distinct strong brown (7.5YR 5/6) mottles; massive; firm, slightly sticky and slightly plastic; common distinct fine manganese stains; few distinct dark grayish brown (10YR 4/2) clay films in pores; gradual smooth boundary.

Jackson Frazier
Plot III

- A1 0-11cm. Very dark grayish brown(10YR 3/2)silty clay loam; moderate, fine granular structure; friable, sticky and plastic; many very fine, many fine and many medium roots; abrupt smooth boundary.
- A2 11-22cm. Very dark gray(10YR 3/1)silty clay loam; many fine distinct reddish brown(2.5YR 4/4) mottles; moderate fine subangular blocky structure; firm, sticky and plastic; common fine roots; clear smooth boundary.
- BA 22-39cm. Very dark gray(10YR 3/1)silty clay; many medium distinct dark reddish brown(2.5YR2.5/2) and many medium distinct dark reddish brown(5YR 3/3), and common fine distinct dark yellowish brown(10YR 4/4)mottles; moderate medium subangular blocky structure; firm, sticky and plastic; common fine roots; many medium black concretions; clear smooth boundary.
- Bw 39-74cm. Very dark gray(10YR 3/1)clay; weak coarse prismatic and weak very coarse prismatic structure; very firm, very sticky and very plastic; few very fine and few medium roots; many medium black concretions; gradual smooth boundary.
- Bss1 74-114cm. Very dark grayish brown(2.5Y 3/2)clay; weak very coarse prismatic structure; very firm, very sticky and plastic; few very fine roots; many common black nodules; common distinct slickensides; vertical cracks 10mm wide; gradual smooth boundary.
- Bss2 114-174cm. Very dark gray(10YR 3/1) 60%, very dark grayish brown(2.5Y 3/2) 40% silty clay; massive; very firm, sticky and plastic; few very fine roots; many medium and many fine black concretions; common distinct slickensides; pressure faces on vertical surfaces with cracks 10mm wide; gradual smooth boundary.

APPENDIX B: TOPOGRAPHIC SURVEY COORDINATES

Origin Point	Destination Point	Magnetic Bearing (degrees)	Distance (meters)
1	2	285	50
2	3	285	50
3	4	285	50
4	5	285	50
5	6	285	50
6	7	285	50
7	8	299	35
8	9	360	50
9	10	360	50
10	11	360	50
11	12	360	50
12	13	360	50
13	14	360	50
11	16	136	50
16	17	75	37
17	18	120	50
18	19	100	50
19	21	86	50
21	22	90	50
22	23	90	50
23	24	90	50
17	25	39	40
25	26	40	40
19	27	38	50
27	28	37	50
28	29	19	50

Origin Point	Destination Point	Magnetic Bearing (degrees)	Distance (meters)
29	30	19	50
22	31	325	40
31	32	50	50
32	33	20	50
33	34	120	25
34	35	355	50
35	36	6	50
28	37	307	50
37	38	210	50
38	39	355	50
39	40	4	50
40	41	296	50
41	42	11	30
42	43	14	19
42	44	250	50
12	45	105	50
45	46	105	30
1	47	37	50
47	48	31	50
48	49	129	50
49	50	135	28
50	51	38	50
51	52	31	50
52	53	45	50
53	54	47	50
54	55	74	40

Origin Point	Destination Point	Magnetic Bearing (degrees)	Distance (meters)
55	56	2	50
56	57	348	40
57	58	93	32
58	59	23	21
59	60	59	50
60	61	5	50
61	62	298	50
62	63	201	50
63	64	223	38
64	65	315	50
65	66	308	25
66	67	300	44
67	68	309	36
68	69	31	50
69	70	31	50
70	71	282	50
71	72	212	50
72	73	269	50
73	74	285	50
74	75	297	50
75	76	255	47
76	77	280	50
77	78	245	50
78	79	315	50
24	80	303	35
80	81	9	30

Origin Point	Destination Point	Magnetic Bearing (degrees)	Distance (meters)
81	82	45	32
82	83	310	34
83	84	12	50
84	85	65	50
85	86	17	29
86	87	303	50
87	88	24	32
88	89	287	50
89	90	5	50
90	91	10	42
91	92	13	36
92	93	302	40
78	94	196	33
94	95	245	20
95	96	210	11
96	97	147	23
97	98	225	19
98	99	191	22
99	100	141	50
100	101	185	36
101	102	185	17
102	103	127	35
103	104	183	26
104	105	149	25
79	106	318	29
106	107	322	30

Origin Point	Destination Point	Magnetic Bearing (degrees)	Distance (meters)
107	108	310	35
108	109	285	23
109	110	173	25
110	111	140	28
111	112	180	15
112	113	134	22
113	114	205	29
114	115	248	19
115	116	145	31
116	117	183	33
117	118	170	50
53	119	174	50
119	120	166	50
120	121	197	50
121	122	213	50
122	123	185	23
123	124	234	27
124	125	184	40
125	126	137	32
126	127	161	45
127	128	152	47
128	129	213	31
129	130	199	37
130	131	166	50
131	73	180	41