SOIL STRATIGRAPHY:

PRINCIPLES, APPLICATIONS TO DIFFERENTIATION AND CORRELATION OF QUATERNARY DEPOSITS AND LANDFORMS, AND APPLICATIONS TO SOIL SCIENCE

A THESIS

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ABSTRACT

Soil stratigraphy comprises the techniques of using "soils", in the sense of weathering profiles, as stratigraphic markers for correlation of Quaternary deposits and landforms. These techniques have been greatly refined during the past decade as a result of research by a comparatively few geologists, including the author; this dissertation is the first comprehensive discussion of this subject.

Modern stratigraphic studies of Quaternary deposits in temperate latitudes invariably demonstrate the occurrence of a sequence of weathering profiles that can be treated as stratigraphic units, called soil-stratigraphic units. The term geosol is proposed as the fundamental soil-stratigraphic unit, to replace the term "soil" used in this sense under the 1961 Code of Stratigraphic Nomenclature. ("Soil" is ambiguous because of its varied meanings.) A geosol is a laterally traceable and mappable layer of distinctly weathered, predominantly mineral matter, formed immediately beneath and generally parallel with the land surface. It maintains a consistent stratigraphic (age) relationship to older and younger deposits with which it is associated, and is defined and used on the basis of this relationship. A unit that evinces distinct surficial weathering but does not meet the relatively strict stratigraphic requirements of a geosol may be called a weathering profile (or a para-geosol). It is also a soil-stratigraphic unit. Geosols are distinct stratigraphic entities; they have certain attributes like, and others unlike, those of both rock-stratigraphic units and unconformities. Geosols may occur either buried by younger deposits (buried geosols) or exposed continuously at the land surface since they were formed (relict geosols). Being laterally traceable, geosols may differ in their physical and chemical

characteristics from one location to another, because of changes in environmental weathering factors, such as climate, vegetation, drainage, and parent material. These lateral changes are called soil facies. Geosols commonly differ from one another in degree of development. Each geosol, however, maintains the same general degree of development relative to the other geosols wherever it occurs in an area, in spite of changes in soil facies. Accordingly, the more strongly developed geosols are excellent stratigraphic markers for differentiation and correlation of Quaternary deposits and landforms within local areas. Criteria are given for identifying and describing geosols and for determining their stratigraphic position and relative age, with examples both of specific situations typically encountered in the field, and also of several modern large-area studies that have used soil-stratigraphy. Identification and correlation problems caused by soil facies and by secondary modification of weathering profiles are discussed.

The physical record indicates that geosols formed during distinct, widely separated intervals of time (from less than 200 to a few thousand years long), in response to infrequent combinations of climatic factors that induced both general land-surface stability (minimal erosion and deposition) and a more accelerated rate of chemical weathering than normal. Appreciably higher-than-normal temperature appears to have been the chief factor that triggered the accelerated weathering. The weathering optima were periodically repeated parts of whole climatic cycles that are manifest in the Quaternary stratigraphic sequences. The thermal optima that induced the weathering optima probably affected the whole earth synchronously, although their climatic effect was proportionately greater at middle and high than at low latitudes. For this reason, geosols probably are nearly time-parallel, and are valid markers also for long-distance correlation (except perhaps at low latitudes), and likewise are valid for defining time-stratigraphic units within the Quaternary.

CHAPTER I

INTRODUCTION

For more than a decade the branch of surficial geology that has come to be known as soil stratigraphy has undergone rapid development because of its increasing use for study of Quaternary deposits and landforms. Soil stratigraphy, as used by Quaternary stratigraphers and geomorphologists, comprises the techniques of using "soils", in the sense of zones of surficial weathering or "weathering profiles", both as stratigraphic markers for differentiating and correlating the Quaternary deposits and landforms of local areas, and also for long-distance correlation of Quaternary units. Geologists using this technique have borrowed heavily from various concepts and methodology of soil science, but have also developed certain other techniques and concepts that are independent of those usual in soil science. The techniques and concepts that are unique to soil stratigraphy have come about largely from field investigations of Quaternary deposits in various lithogenetic terrains (glacial, pluvial lacustrine, alluvial, etc.) by a relatively few geologists in the U. S. and a few foreign countries, notably Australia. This specialty has developed so rapidly that publication of its concepts and applications has been limited to the reports of these field investigations, and generally these reports give only the details that are pertinent to the area in question. No comprehensive survey of the whole subject of soil stratigraphy has yet been published. The present dissertation attempts not only to compile and integrate the information of this subject in U. S. and foreign literature, but also to add considerable new material based on my own experience during the past 16 years in using and developing the technique in stratigraphic work with the U. S. Geological Survey in various parts of the western U. S. It is hoped that this dissertation will serve as a manual for the geologist (and perhaps the soil scientist) seeking to learn how to use this valuable technique in his own work.

Previous work

Literature on weathering profiles as stratigraphic units is relatively voluminous and dates back to before the turn of the century; however, papers that deal with the more sophisticated concepts and practices of soil stratigraphy are comparatively few and have all been published within the last 15 years. The earlier studies were concerned with buried soils, whose "fossil" state and stratigraphic relations are relatively obvious. Recognition of the antiquity of many soils that are now exposed at the land surface has come only within recent years with improved techniques of soil stratigraphy. In the U. S., until little more than a decade ago, geologists were virtually alone in the study of ancient soils; only recently have soil scientists entered this field of investigation.

The first identification of weathering profiles as stratigraphic units was by the early geologists in the midcontinent region (e.g., Worthen, 1866, 1873; McGee, 1891; Leverett, 1898a, 1898b, 1898c, 1899). These workers observed three strong weathering profiles, which they named the Yarmouth, Sangamon, and Peorian soils, and recognized that they record interglacial intervals between the major glacial stages. Subsequently, many geologists recognized and used weathering profiles in their studies, primarily as an aid for local correlation, but also in some cases for evidence on the climatic and geomorphic history of an area. The more important studies in the midcontinent glaciated region are those of Leverett, 1902, 1909; Kay, 1916, 1931; Kay and Pearce, 1920; Kay and Apfel, 1929; Leighton, 1923, 1926, 1931; Leighton and MacClintock, 1930; Leighton and Willman, 1950; MacClintock, 1933; Kay and Graham, 1943; Fisk, 1951; Frye, Willman, and Glass, 1962; Willman, Glass, and Frye, 1963). From this geographic base, stratigraphic use of weathering profiles has spread through most of the United States: to the Great Plains (for example,by Lugn, 1935, 1941; Albritton and Bryan, 1939; Bryan and Albritton, 1943; Hobbs, 1945; Schultz and Stout, 1945; Schultz, Lueninghoener, and Frankforter, 1951; Condra, Reed, and Gordon, 1947; Frye and Fent, 1947; Frye, Swineford, and Leonard, 1948; Frye, 1949a, 1949b, 1951; Frye and Leonard, 1949, 1952, 1954, 1957; Hunt, 1954; Miller and Scott, 1955; Malde, 1955; Handy and Davidson, 1956; and Scott, 1963); to the Rocky Mountain region (for example, by Hunt, 1948, 1953a; Hunt, Creamer, and Fahey 1949; Hunt and Sokoloff, 1950; Moss, 1951; Holmes, 1951; Holmes and Moss, 1955; Leopold and Snyder, 1951; Leopold and Miller, 1954; and Richmond, 1949, 1954, 1960, 1961, 1962a, 1962b, and in press); to the Basin-and-Range Province (notably by Hunt, 1953b; Eardley, Gvosdetsky, and Marsell, 1957; Morrison, 1952a, 1952b, 1961a, 1961b, 1961c, 1961d, and in press, a and b; and Morrison and Frye, in preparation); and even to the southeastern states (e.g., Eargle, 1940; Parizek and Woodruff, 1957; and Hunt and Hunt, 1957).

The first formal proposals that soils (weathering profiles) be classed as stratigraphic units, also giving some of the stratigraphic attributes of weathering profiles used as stratigraphic markers, were by Richmond and Frye (1957) and Richmond (1959).

Investigations by soil scientists of weathering profiles from the stratigraphic viewpoint have dealt mainly with buried profiles. The principal papers are those of Sturgis and McMichael, 1939; Thornbury, 1940; Simonson, 1941, 1954; Wascher, Humbert, and Cady, 1947; Krusekopf, 1948; Thorp, 1949; Scholtes, Ruhe, and Riecken, 1951; Thorp, Johnson, and Reed, 1951; Ruhe, 1952, 1956, 1962; and Hogan and Beatty, 1963.

From the entire group of papers listed above, the papers that have made the most important contributions to development of concepts and practice of modern soil stratigraphy are those of Bryan and Albritton, 1943; Thorp, 1949; Hunt and Sokoloff, 1950; Thorp, Johnson, and Reed, 1951;

Richmond and Frye, 1957; Richmond, 1959, 1962a; Ruhe, 1962; Scott, 1963; and Morrison, in press, a and b).

Abroad, the pioneer stratigraphic observations on weathering profiles appear to have been those made on buried soils in the steppes of western Russia by Vysotskii (1901, cited by Joffe, 1949). Further important studies in western Russia include those of Nabokikh (1914), Polynov (1927), Florov (1928), and Glinka (1932) (all cited by Joffe, 1949), and also by Krokos (1926, 1927) (cited by Thorp, Johnson, and Reed, 1951), Nabokikh (1916), Prasolov and Sokolov (1927), and Ponomarev and Sedletzkii (1940) (cited by Hunt and Sokoloff, 1950). Notable studies, also mainly on buried soils, also have been made in Germany (e.g., Schonhals, 1950; Brunnacker, 1957); in Mexico (K. Bryan, 1948); in China (Pendleton et al., 1932, 1935; Thorp, 1935a, 1935b, 1936; Movius, 1949); in Java (Mohr, 1944; Movius, 1949); in India (Movius, 1949; Raychaudhuri and Sen, 1952); and in Australia (W. Bryan, 1939; Northcote, 1946, 1951; Walker, 1962a, 1962b; Butler, 1958a, 1958b, 1959, 1960, 1961; van Dijck, 1958, 1959; and Jessup, 1960a, 1960b, 1961).

What is a soil?

Soil, like many common old words, has a multiplicity of meanings. In its traditional and most common meaning among farmers, livestock growers, and foresters, it is virtually any unconsolidated surficial material (mineral and/or organic) capable of growing vegetation; in this sense its thickness is determined by the depth of rooting of plants growing on it. Another meaning is earth darkened by organic matter. Most laymen and even specialists such as soil-mechanics engineers and many geologists also use the term soil very loosely, as a general term for any unconsolidated surficial material, whether or not it meets the pedologic definition of soil by having developed soil horizons. To some construction engineers, soil is any earth material that is sufficiently poorly indurated to be worked with power excavating equipment, without requiring blasting. The attitude of many geologists toward soil is stated candidly by Richthofen (1888, cited by Joffe, 1949, p. 17) "Soil is a loose surface formation, a kind of pathological condition of the native rock", and by De Beaumont (1845, cited by Tolman, 1937, p. 122) "The soil often hinders geological research by covering the rocks in place".

Definitions by soil scientists have tended to stress the biologic agencies manifest in soil. Marbut (1951), for example states that "the soil is that layer of the earth's crust lying within reach of those forces which influence, control, and develop organic life". More recently soil is defined (Soil Survey Staff, 1960, p. 1) as:

> "........the collection of natural bodies on the earth's surface, containing living matter, and supporting or capable of supporting plants.......Soil includes all horizons differing from the underlying rock material as a result of interactions between climate, living organisms, parent

materials, and relief.........commonly, soil grades at its lower margin to hard rock or to earthy materials essentially devoid of roots. The lower limit of soil therefore is normally the lower limit of the common rooting of the native perennial plants....."

Soil scientists also have coined more specific terms analogous to soil, such as soil profile, solum, pedon and paleosol. These terms are discussed in the next chapter.

Soil-stratigraphic units

Recently the small group of geologists that uses soils as stratigraphic markers for study of Quaternary deposits and landforms has begun to use the term soil in a much more restricted sense, to mean a profile of surficial weathering, which generally is called a "weathering profile".[1/] This usage has been formalized in the recently revised Code of Stratigraphic Nomenclature (American Commission on Stratigraphic Nomenclature, 1961, p. 654-655). In the revised Stratigraphic Code the concepts of soil stratigraphy were recognized by setting up a new classification called soil-stratigraphic units. A soil-stratigraphic unit is defined

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1/ A weathering profile can be defined as a layer of earth material, that is now or was in the past exposed at the earth's surface, and shows criteria of pedochemical weathering sufficiently strong to be discernible in the field by competent observers. Pedochemical weathering is evinced most commonly by: (1) Deposition of layer silicates (clay minerals) and sesquioxides (of Fe, Al, and Mn) in the B horizon, as a result of hydrolysis and oxidation of the mineral parent material, and of translocation of colloidal clay. Increase in clay-sesquioxide content typically causes development of thicker and more numerous clay skins, higher grades of soil structure, more plastic and harder consistence, and redder hue and deeper chroma. (2) Leaching of calcium carbonate (and other salts such as gypsum, and in some cases, silica) from the upper part of the soil profile, with or without redeposition (concentration) of these compounds in the lower part of the profile. The minimum degree of pedochemical weathering for a weathering profile generally is the minimum requirement for a cambic or "color B" horizon (see p. 18).

as a soil (weathering profile) with physical features and stratigraphic relations that permit its consistent recognition and mapping as a stratigraphic unit; i.e., it is a definite stratigraphic entity that is laterally traceable and mappable. The single rank of the soil-stratigraphic classification, as presently established, is the soil. Soil-stratigraphic units (soils in this restricted sense) are characterized as "the products of surficial weathering and of the action of organisms at a later time and under ecologic conditions independent of those that prevailed while the parent rocks were formed". Provision of the separate soil-stratigraphic classification was made in recognition of important differences between soils and both rock-stratigraphic units and pedologic units (see chapters 3 and 7).

Proposal of the term geosol as the fundamental soil-stratigraphic unit

Although the term "soil" is given as the fundamental and only soil-stratigraphic unit in the revised Stratigraphic Code, this common name is so ambiguous by reason of its many meanings that the writer believes it is desirable to provide a new term that specifically identifies the type of soils that are used as fundamental soil-stratigraphic units; in other words, a term that identifies them as geologic, rather than pedologic, soil-mechanics, or other soil units. The term geosol is proposed for soils defined and used in this sense. No existing term, such as soil profile, paleosol, etc., connotes this meaning of a unit that is a definite stratigraphic entity.

Definition of geosol.--The term geosol is proposed as the fundamental soil-stratigraphic unit to replace the term "soil" used in this sense under the 1961 Code of Stratigraphic Nomenclature. A geosol is a laterally traceable and mappable layer of distinctly weathered, predominantly mineral material, formed immediately beneath and generally parallel with the land surface, that maintains a consistent stratigraphic relationship to the deposits with which it is associated (i.e., a consistent age relationship to older and younger deposits), and is defined and used on the basis of this relationship within reasonably exact limits. A geosol must evince sufficient alteration by surficial weathering-pedogenic processes to be discernible in the field; particularly, it must manifest at least the minimum requirements for a cambic[1/] (or a "color" B horizon (see Soil Survey Staff, 1960, p. 49-50). A profile that consists merely of an A horizon directly over a C horizon (without a B horizon), or solely of a C horizon, does not qualify. Thus, a geosol is an alteration deposit, formed on and commonly from underlying rock-stratigraphic units. It is considered to be stratigraphically separate from the underlying rock-stratigraphic units, inasmuch as these may be of diverse age and they are invariably older than the geosol itself and deposited under

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[1/] A cambic horizon (from Latin Cambiare, to change) is one of the kinds of subsurface soil horizons under the new U. S. (Soil Survey Staff, 1960) soil classification system. It occurs only in textures finer than loamy very fine sand; it lies in the position of the B horizon; it contains feldspars, micas, or other weatherable minerals; it generally shows little of the original rock structure, but commonly has soil structure; it shows at least some evidence of alteration (weathering), such as formation of clay, loss of "combined" or "lattice" iron, and/or redistribution of carbonates. The minimum requirement for a cambic or "color" horizon is a higher chroma, or redder hue, than the C horizon; it is not necessary to have clay skins or any appreciable increase in clay content.

different environmental conditions. A geosol may occur either at, or at any depth below the earth's surface.

Each geosol is given a local geographic name that distinguishes it from other stratigraphic units. These names are independent of the geographic names given by soil scientists to soil series units.

A geosol is distinct from pedologic units; it may include one or more pedologic units (or parts of these units), but not all pedologic units are geosols, and commonly the lateral and vertical boundaries of the two types of units are different. The lower boundary of a geosol is either the base of the solum (base of the B horizon) if no Cca horizon is present, or the base of the normal Cca horizon if this horizon is present. This limitation excludes still deeper weathering zones that are detectable in some weathering profiles (p. 28).

Not-quite geosols.--The stratigraphic requirements for defining a geosol are fairly rigorous. In some cases one encounters a unit that obviously is a weathering profile, but whose stratigraphic position and relative age is known only approximately. Such a unit should not be called a geosol, but by a term that identifies it as a not-quite geosol, such as "para-geosol", "weathering profile", or (less desirably) "soil". A para-geosol (weathering profile) is itself a soil-stratigraphic unit; it has all the basic stratigraphic attributes and modes of stratigraphic occurrence that geosols have, and it should be defined like a geosol insofar as possible. The problem of terminology of geosols vs. para-geosols is discussed further in chapter 3.

The term geosol commonly is used alone in this paper for reasons of brevity, although in most cases the discussion pertains equally to para-geosols. In the few cases where distinction must be made between the terms, the appropriate one is used.

Age of a geosol (and para-geosol)

A geosol (or para-geosol), being a new deposit transformed by weathering from its various parent material units, is considered to date from the soil-forming interval during which it developed its distinctive profile characteristics. It is invariably younger than the age of the youngest parent material unit or youngest geomorphic (erosional or depositional) surface on which it developed, and it is older than the oldest deposit that overlies it. Detailed stratigraphic study of the surficial deposits in a comparatively large area generally is needed in order to bracket the geologic age of a weathering profile with sufficient precision for it to qualify as a geosol. The age relations of a weathering profile may be determined in relation to a sequence of geomorphic surfaces. However, such a determination generally is much less specific than one made from detailed stratigraphic study, and may not be adequate for formal definition of a geosol. General techniques for determining the age of a geosol are given in chapter 4.

Naming of geosols and weathering profiles

As pointed out above, a geosol should be given a geographic name that distinguishes it from other stratigraphic units (rock-, soil-, time-, bio-, and geologic-climatic). It should not be named after another stratigraphic unit within which it is intercalated or on which it rests (e.g., if a geosol is interstratified with or rests on the Podunk Formation, it should not be called the Podunk Geosol). If geosols are not given formal names of their own, they may be named informally in terms of their age relations to rock-stratigraphic (or geologic-climatic) units that bracket them in age, or merely, for brevity, as post-dating the youngest underlying rock-stratigraphic or geologic-climatic unit [(e.g., post-Bull Lake geosol (in reference to the Bull Lake Glaciation); post-Alpine geosol (in reference to the

Alpine Formation)]. If the less specific terms weathering profile or soil are used, the unit can be named informally after the youngest geomorphic surface on which it rests (e.g., post-Broadway terrace soil, post-intermediate pediment weathering profile).

Some geologists have named soils after time-stratigraphic (or geologic age or geologic-climatic) units (e.g., the "Wisconsin" and "Recent" soils of Hunt (1954), Malde (1955), and Scott (1963) in the vicinity of Denver, Colorado). This practice is inadmissable for geosols and is not recommended even if the units are called weathering profiles or soils. Aside from the undesirability of mixing soil-stratigraphic and other stratigraphic names, the stratigraphic and age relations of the soil unit are likely to be ambiguous, particularly if the time-rock unit is large. For example, at least half a dozen weathering profiles have been identified intercalated with deposits that are correlated with the Wisconsinan Stage (Glaciation), notably in the Lake Lahontan and Lake Bonneville areas. To avoid any possible ambiguity it is better either to give the soil unit a formal name, or to give as specific an informal name as possible (such as lower intra-Bull Lake weathering profile, lower intra-Eetza soil, upper pre-Lake Bonneville geosol).

CHAPTER II

ELEMENTS OF SOIL DESCRIPTION AND CLASSIFICATION FOR GEOLOGISTS

To many geologists, soils are bodies of little importance or interest; indeed, they are annoying hindrances to their work wherever they compound Mother Nature's inscrutability by concealing the bedrock. The geologist who has the patience to study soils closely begins to recognize that soils are independent natural bodies with distinctive morphologies. It commonly comes as a surprise to find how infinitely variable soils can be, how many properties they have and how difficult their total aspect is to describe in words. Soil scientists have tried to cope with this problem by developing a very specific terminology for describing soils. Some elements of this terminology are akin to terms used by geologists for describing lithologic characteristics of sedimentary deposits, but most elements are diferent, inasmuch as soils have many properties that are rarely, if ever, possessed by unweathered sedimentary deposits, and also soils are described in much more detail than is customary for sedimentary deposits. The soil-scientists' descriptive terminology is admirably suited for describing the essential characteristics of soils fully, unambiguously and objectively, particularly the characteristics that are observable in the field. It is desirable that geologists adopt this system instead of trying to describe soils by means of ordinary geologic descriptive terms (which commonly are much less specific and exact), or trying to develop another descriptive system of their own.

In order to do this, it obviously is necessary for the geologist to become familiar with the essentials of the soil-descriptive terminology. The terminology that is generally used by soil scientists in the U. S. is that adopted by the Soil Survey staff of the Soil Conservation Service,

U. S. Dept. of Agriculture, and published in the Soil Survey Manual (Soil Survey Staff, 1951). Definitions of certain terms have recently been modified (particularly soil horizon terminology); the summary below gives the usage as of 1964.

Soil-descriptive terminology

The soil profile and related terms

Soil scientists use the term soil profile for the whole aggregate of the various soil horizons measured vertically in section. The Soil Conservation Service (1962, p. 173) defines a soil profile as "........the collection of all the genetic horizons, the natural organic layers on the surface, and the parent material or other layers beneath the solum that influence the genesis and behavior of the soil". In practice, this term is used for any kind of surficial profile, both those that have well developed genetic soil horizons (such as B and possibly Cca horizons), and for those that are essentially unweathered throughout, such as those with an organic A horizon directly underlain by an unmodified C horizon, or even those with just an unmodified C horizon from top to bottom. Thus, the term soil profile is not synonymous with weathering profile and should not be used in this sense.

Soil profiles vary almost infinitely. They range in thickness from mere films to scores of feet. They vary widely in the degree to which genetic (weathering) horizons are expressed. On very young geologic deposits, such as modern alluvium or eolian sand, genetic horizons may be completely absent. The earliest stages of weathering and genetic changes can be detected only by laboratory study of samples; later as weathering progresses the genetic changes can be seen with increasing clarity in the field.

A soil profile that shows evidence of removal of its upper portion by geologic erosion is called a <u>truncated</u> profile (or truncated soil) by soil scientists.

The <u>solum</u> is defined (Soil Survey Staff, 1951, p. 183) as ".......the genetic soil developed by soil-building forces. In normal soils......(it) includes the A and B horizons, or the upper part of the profile above the parent material". The solum usually is not considered to include the C (or Cca) horizon (fig. 2.1). However, in some intrazonal (immature) soils the lower limit of the solum can only be set arbitrarily; in such cases it usually is put at the lower limit of plant roots--which commonly means that it includes C material. (Obviously there is uncertainty among soil scientists as to the definition of solum.) Further, in deep lateritic soils the base of the solum is controversial; although the upper part of the B horizon obviously is part of the solum, the tendency is to exclude the lower part because it is "far removed from the influence of organisms".

<u>Soil horizons</u>

A soil horizon is defined (Soil Survey Staff, 1960, p. 24; Soil Cons. Service, 1962, p. 173) as a layer within a soil that is approximately parallel to the soil surface and that has properties that are produced by soil-forming processes, that are unlike those of adjoining layers. Horizons therefore are identified partly by their own characteristics and partly by the properties of underlying and overlying horizons. A soil horizon is differentiated from those adjacent to it mainly by characteristics that can be seen or measured in the field, such as color, structure, texture, consistence, and the presence or absence of carbonates. Laboratory data sometimes may supplement the designation and detailed characterization of horizons.

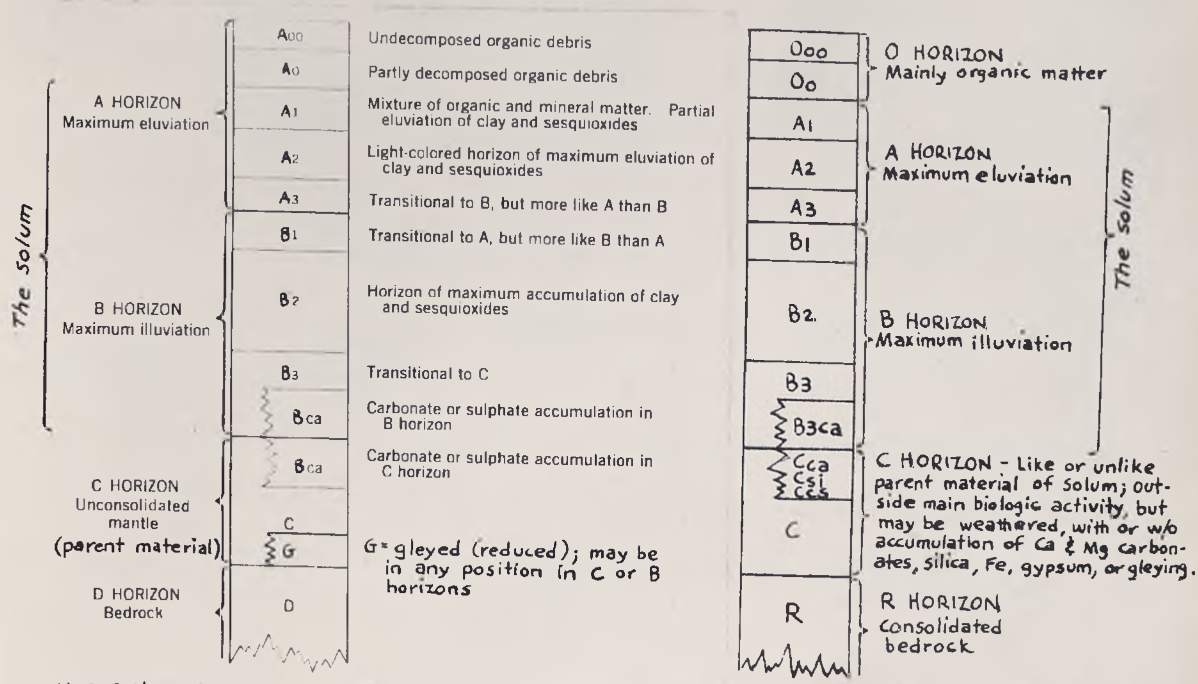
Each soil horizon has chemical and physical characteristics that distinguish it from other soil horizons in a soil profile and from the underlying parent material. These characteristics generally become more sharply differentiated with increasing development of the soil. However, besides horizonation resulting from pedogenic processes some soil profiles have layering inherited from stratified parent material. Thus, many soils have profiles whose properties partly result from soil-forming processes, partly are inherited from stratified parent material, and even partly are due to geologic processes accompanying soil formation (for example, a soil may be gradually covered with volcanic ash, loess, eolian sand, or alluvium without seriously injuring the vegetation, so that the surface horizon becomes thickened).

In describing a soil profile, soil scientists usually locate the boundaries between horizons, measure their depth below the land surface, and study the profile as a whole before describing and naming the individual horizons.

Fig. 2.1 shows how soil horizons are defined and differentiated by the Soil Survey Staff, U. S. Soil Conservation Service (1951, 1962). This organization recently has revised its mode of soil horizon designation and differentiation, hence both the former and current schemes are shown in fig. 2.1,A. The figures are schematic and not all of the horizons and subhorizons shown are found in a single soil profile.

The so-called master soil horizons are designated by capital letter symbols: formerly A, B, C, G, D; now changed to O, A, B, C, R. They indicate major kinds of departures from parent material.

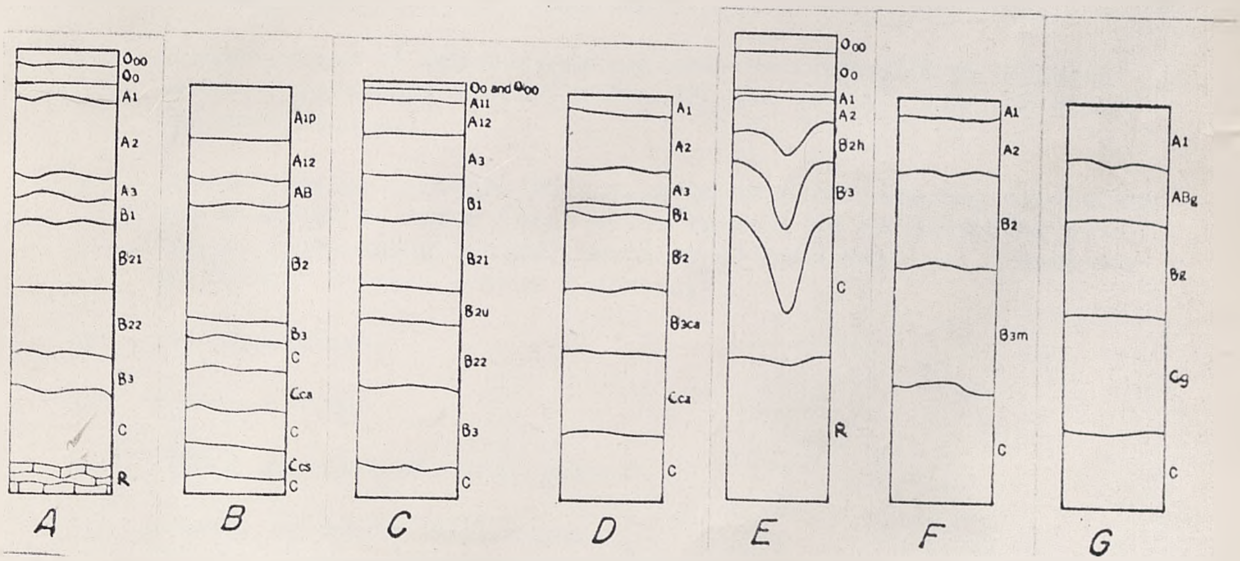
Subdivisions of the master horizons are shown by placing an Arabic number (1, 2, or 3) after the capital letter, thus: A1, A2, B1, B2, B3, etc. Generally, the A1 and B1 symbols indicate subhorizons transitional from the



U. S. Soil-horizon classification prior to 1962.

U. S. Soil-horizon classification since 1962.

A. Possible horizons in soil profiles.



B. Hypothetical profiles to illustrate a few horizon designations: A, Gray-Brown Podzolic; B, Chernozem (cultivated); C, Latosol; D, solodized-Solonetz; E, Podzol; F, Planosol; G, Humic Gley (Wiesenboden).

Fig. 2.1. Soil horizon designation (modified after Soil Survey Staff,1951).

overlying horizon, and A3 and B3, transitional into the underlying C horizon. The symbol B2 indicates that part of the B horizon that is of a nature not transitional either to A or to C. Even if both B1 and B3 are absent and if the B horizon of a given profile is not subdivided, the symbol B2 and not B is used.

The next smaller degree of subdivision is indicated by additional Arabic numbers (1, 2, 3, etc.) placed to the right of the numbers that indicate the main subdivisions; eg., A11, A12, B21, B22, and B23.

Certain kinds of special departures from the master horizons are indicated by lower case letter suffixes to the capital and numeral letter symbols, thus: B3ca, Cca. These symbols, as currently used, are as follows:

b: buried soil horizon.

ca: accumulation of carbonates of alkaline earths, commonly of calcium.

cs: accumulation of calcium sulfate.

cn: accumulations of concretions or hard nonconcretionary nodules enriched in sesquioxides (of Fe, Al, Mn, or Ti), with or without phosphorous.

g: strong gleying (intense reduction of iron or reducing conditions due to stagnant water (high water table), as evidenced by base colors approaching neutral, with or without mottles). Replaces the former capital letter G symbol.

h: illuvial humus.

ir: illuvial iron.

m: strong cementation, induration.

p: plowing or other disturbance.

sa: accumulation of salts more soluble than calcium sulfate.

si: cementation by siliceous material, soluble in alkali, as defined for duripans (applied only to C horizon).

t: illuvial clay (applied to B horizon, and commonly called a "textural B" horizon).

x: fragipan character.

A lithologic discontinuity is a textural or mineralogic change that indicates an appreciable (geologic) difference in the parent material, as contrasted to horizon differences resulting from pedogenesis. Important lithologic discontinuities, either within or below the solum, are shown by Roman numerals prefixed to the master horizon symbol. The first, or uppermost, material is not numbered, for the Roman numeral I is understood; the second contrasting material is numbered II and others encountered below are numbered III, IV, and so on. Thus a sequence from the surface downward might be A2, B1, IIB2, IIB3, IIC1, IIIC2.

Unfortunately, soil-horizon designation as currently practiced by most U. S. soil scientists stops short of the lower limit of detectable weathering changes in the more mature weathering profiles. Jackson and Sherman (1953) point out that the geochemical weathering profile may be deeper than the solum or soil profile as usually defined, and cite several soil scientists who prefer to extend their definition of soil to include <u>all</u> weathered zones.

A common example of weathering changes that extend considerably below the soil profile as usually designated can be seen in pedalfer[1/] weathering profiles developed in calcareous tills or other initially calcareous parent materials. Beneath the solum of these profiles commonly there is a zone that is leached of primary carbonates. This zone may be strongly to weakly oxidized, but it lacks soil structure and is considered to be C horizon (parent material) by soil scientists. Beneath this zone

[1/] See p.41 for definition of pedalfer.

generally lies a calcareous zone, containing primary carbonates, but still somewhat oxidized. In some areas a subzone occurs at the top of this zone, with primary calcite leached out but with dolomite remaining. In other areas a subzone of calcium carbonate concentration occurs at or near the top of the oxidized-calcareous zone. Such relations are widespread in the continental drift sheets of Midwestern and Northeastern North America (Leighton and MacClintock, 1930; MacClintock and Apfel, 1944; Frye, Willman and Glass, 1960), and also occur in the Cordilleran drifts (e. g., Morrison, in press, b). Other types of weathering profiles, notably latosols and various pedocals with thick caliche (petrocalcic) horizons, commonly extend considerably below the solum and Cca horizon (if present) as usually measured (Soil Survey Staff, 1951, p. 183-184).

In a study of weathering profiles of pre-Wisconsinan age in their entirety in Illinois, the geologists Frye, Willman, and Glass (1960) developed the following modification and extension of the existing standard pedologic terminology: (Note that the various horizons are called "zones", in order to distinguish them from pedologic usage.)

Table 2.1

Weathering profile terminology for continental glacial and associated deposits in Illinois, after Frye, Willman, and Glass, 1960, p. 2-3

A zone	=	The A-horizon with subdivisions of A1, A2, and A3, of standard pedologic terminology.
B zone	=	B-horizon, with subdivisions of B1, B2, and B3, of standard pedologic terminology. Maximum clay enrichment (illuviation) and development of soil structure; commonly oxidized to red or brown, with Mn-Fe pellets.
BG zone	=	Gleyed B-horizon, all or part of which has developed under reducing (water-logged) conditions, commonly as a secondary modification after a primary B horizon. Gray; clay accumulation, but limited soil structure; may have Mn-Fe pellets.
G zone	=	Accretion-gley zone (as defined by Frye, Willman, and Glass, 1960). Clayey fine-textured colluvium, transported laterally from adjacent gentle slopes and deposited in small water-logged depressions, under a reducing environment. Gray; may have some organic matter; locally indistinctly bedded; soil structure generally absent.
C zone	=	Weathered parent material, generally next below B zone, but in places below an A-, BG-, or G-zone. Subdivided into a CL zone and a CC zone where the parent material was initially calcareous.
CL zone	=	Leached C zone. Zone below the B zone that is leached of primary carbonates; may be strongly to weakly oxidized but lacks soil structure.
CC zone	=	Calcareous C zone. Below the CL zone (where present); contains primary carbonates; somewhat oxidized; has structure of the parent material. In places has an upper subzone from which calcite is leached but dolomite still remains.

General terms for degree of development of geosols

The intensity of development of the various diagnostic characteristics listed in the next section determines what is called by geologists the "degree of development" of a geosol. If geosols are to be compared for local rock-stratigraphic or geomorphic differentiation and correlation, however, it is important to note that any of these characteristics may vary between different facies (see chapter 3) of a given geosol (with changes in parent material, climate, relief, vegetation, and drainage), and that it is necessary to compare similar facies of a sequence of geosols (precautions to be taken are discussed in chapter 4).

If the B horizons of two geosols are compared on this basis, one may be consistently thicker, more clayey, redder, more blocky or prismatic, more plastic and harder, and more acid than the other B horizon. Likewise, the Cca horizon (if present) of the first geosol may be consistently thicker and more enriched in calcium carbonate than the Cca horizon of the second geosol. The first geosol is said to be better developed than the second. The following general terms are used to describe relative general degrees of development between geosols: <u>weakly developed</u>, <u>moderately developed</u>, <u>strongly developed</u>, and <u>very strongly developed</u>.

<u>Diagnostic characteristics of soil horizons</u>

Certain characteristics of soil horizons are readily identifiable in the field and are used by soil scientists and geologists alike both to differentiate various horizons and subhorizons and to identify whether or not the unit is a weathering provile (geosol), and if so, its degree of (weathering) development. The chief diagnostic characteristics of soil horizons are: thickness, boundaries, texture, color, structure, consistence, reaction, as well as any concentrations of calcium carbonate, gypsum, other salts, silica, and oxides of iron and/or manganese. The following section summarizes current definitions and practice in designating these characteristics among U. S. soil scientists.

Terms used in describing soil horizons

Boundaries.--Boundaries between soil horizons are described by the following thickness terms:

abrupt: less than 1 inch wide

clear: 1 to $2\frac{1}{2}$ inches wide

gradual: $2\frac{1}{2}$ to 5 inches wide

diffuse: more than 5 inches wide.

Their lateral regularity is described by these terms:

smooth: nearly a plane

wavy or undulating: pockets are wider than their depth

irregular: pockets are deeper than their width

broken: parts of the horizon are unconnected with other parts.

Color.--Soil colors are described by the Munsell system of soil-color notation and nomenclature, in terms of hue, value, and chroma, as measured by use of Munsell Soil Color Charts.[1/] It should be noted that some of the soil-color names differ from standard Munsell color names, although the Munsell notation symbols are the same. Soil colors commonly are described separately for three standard moisture states: dry, moist, and wet; the moisture state always should be given.

Texture.--Soil-texture names differ from geologic (sedimentary size-grade) terminology. Soil-textural classes refer to the proportions of clay, silt, and sand below 2 millimeters in diameter. Most of the textural classes are mixtures of these sizes in various proportions, inasmuch as soils commonly are less well sorted than many sedimentary deposits. The principal textural classes, based on the relative proportion of individual size groups of soil particles are:

1/ Munsell soil color charts are available from the Munsell Color Company, Baltimore 2, Md.

Table 2.2

Principal soil textural classes

General soil-textural terms		Basic soil-textural class names
Sandy soils	Coarse-textured soils	Sand Loamy sand
Loamy soils	Moderately coarse-textured soils	Sandy loam
	Medium textured soils	Loam Silt loam Silt
	Moderately fine-textured soils	Clay loam Sandy clay loam Silty clay loam
Clayey soils	Fine-textured soils	Sandy clay Silty clay Clay

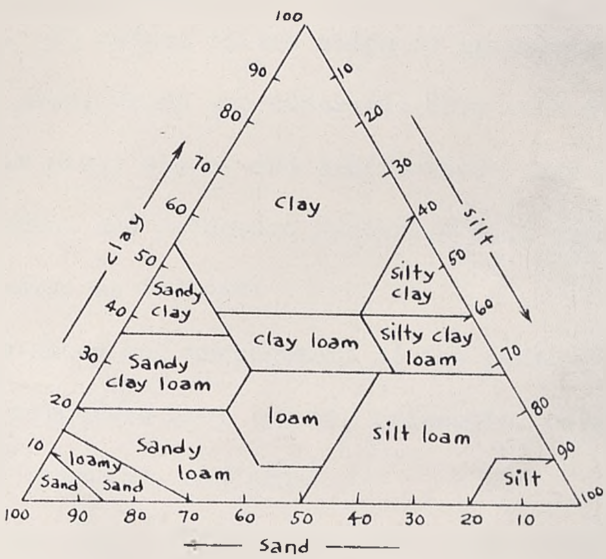


Fig. 2.2. Diagram showing the percentages of clay (below 0.002 mm.), silt (0.002 to 0.05 mm.), and sand (0.05 to 2.0 mm.) in the basic soil textural classes. (Modified from Soil Survey Staff, 1951, p. 209.)

Soil scientists use the following criteria for field identification of soil textural classes:

Loamy sand:	harsh, gritty feel; very slight tendency of the moist soil to stick together when pressed.
Sandy loam:	definitely gritty; when moist the soil may be pressed into a soft mass, and when moistened to its maximum sticky point it is not perceptibly sticky.
Fine sandy loam:	mellow and only moderately gritty feel; when moist, the soil may be pressed into a firm mass. At its maximum sticky point it is very slightly sticky.
Loam:	mellow, moderately smooth feel; moist soil may be rolled into firm rods; slightly sticky.
Silt loam:	smooth and "floury" feel; moist soil rather sticky; readily rolled into firm, slender rods.
Clay loam:	very smooth "slippery" feel; definitely sticky when moist; easily modeled into any shape.

Structure.--Soil structure is one of the most diagnostic properties of a soil horizon. It refers to the shape of aggregates of primary soil particles, called peds, which are separated from each other by surfaces of weakness. Their size, shape, and distinctness vary widely with different parent material and pedogenic factors. Soil structure is described in three ways, as follows:

A) Type.--the shape and arrangement of the peds.

The types of soil structure are: platy, prismatic, columnar, blocky, subangular blocky, granular, and crumb.

B) Class.--dimensional class within any given type.

C) Grade.--degree of development of the structural type.

Table 2.3 gives the types and classes of soil structure as defined by the U. S. Soil Conservation Service.

TABLE 2.3—*Types and classes of soil structure*

[After Soil Survey Staff, 1951, p. 228]

Class	Type (shape and arrangement of peds)						
	Platelike aggregates arranged in a horizontal plane; faces mostly horizontal	Prismlike aggregates with horizontal dimensions considerably less than vertical. Vertical faces well defined; vertices angular		Blocklike, polyhedral, or spheroidal aggregates, with three dimensions of the same order of magnitude			
				Plane or curved surfaces that are casts or molds of faces of adjacent peds		Plane or curved surfaces that have slight or no accommodation to faces of adjacent peds	
		Caps flat	Caps rounded	Faces flat; vertices sharply angular	Faces rounded and flat; many rounded vertices	Relatively nonporous peds	Porous peds
	Platy	Prismatic	Columnar	Angular blocky	Subangular blocky	Granular	Crumb
Very fine or very thin.	Very thin platy; < 1 mm.	Very fine prismatic; < 10 mm.	Very fine columnar; < 10 mm.	Very fine angular blocky; < 5 mm.	Very fine subangular blocky; < 5 mm.	Very fine granular; < 1 mm.	Very fine crumb; < 1 mm.
Fine or thin	Thin platy; 1 to 2 mm.	Fine prismatic; 10 to 20 mm.	Fine columnar; 10 to 20 mm.	Fine angular blocky; 5 to 10 mm.	Fine subangular blocky; 5 to 10 mm.	Fine granular; 1 to 2 mm.	Fine crumb; 1 to 2 mm.
Medium	Medium platy; 2 to 5 mm.	Medium prismatic; 20 to 50 mm.	Medium columnar; 20 to 50 mm.	Medium angular blocky; 10 to 20 mm.	Medium subangular blocky; 10 to 20 mm.	Medium granular; 2 to 5 mm.	Medium crumb; 2 to 5 mm.
Coarse or thick	Thick platy; 5 to 10 mm.	Coarse prismatic; 50 to 100 mm.	Coarse columnar; 50 to 100 mm.	Coarse angular blocky; 20 to 50 mm.	Coarse subangular blocky; 20 to 50 mm.	Coarse granular; 5 to 10 mm.	
Very coarse or very thick.	Very thick platy; >10 mm.	Very coarse prismatic; >100 mm.	Very coarse columnar; >100 mm.	Very coarse angular blocky; >50 mm.	Very coarse subangular blocky; >50 mm.	Very coarse granular; >10 mm.	

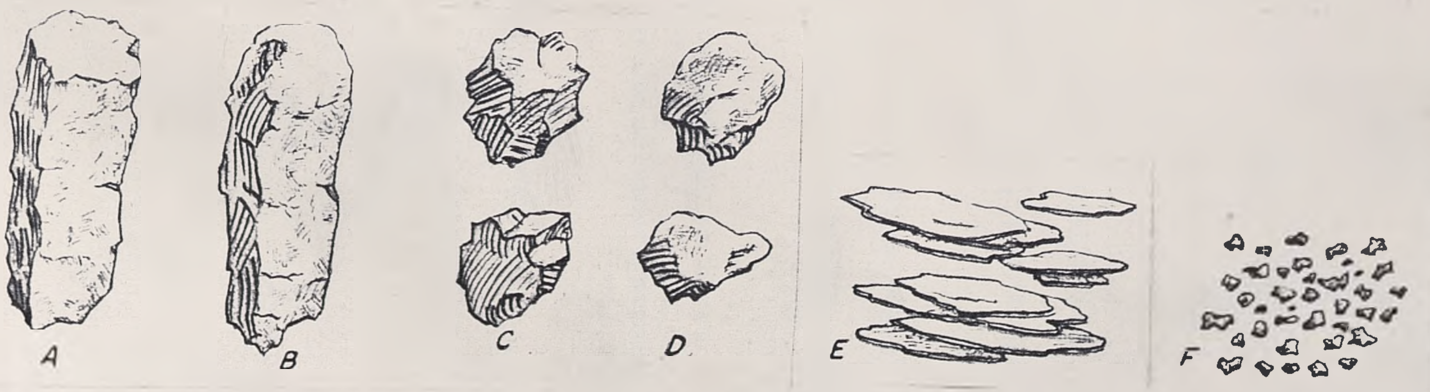


Fig. 2.3. Drawings illustrating types of soil structure: A, prismatic; B, columnar; C, angular blocky; D, subangular blocky; E, platy; and F, granular. (After Soil Survey Staff, 1951.)

Terms for grade of soil structure are as follows:

Structureless:[1/] No observable aggregation. Or:

Massive, if coherent, or

Single-grain, if incoherent.

Weak: Poorly-formed indistinct peds. When disturbed, these break into a mixture of a few entire peds, many broken peds, and much unaggregated material.

Moderate: Well-formed, distinct, and moderately durable peds, which when disturbed break down into many distinct peds, some broken ones, and little unaggregated material.

Strong: Well-formed, distinct, and durable peds, which resist displacement and break down into many entire peds, a few broken ones, and little unaggregated material.

Soil-structure type, class, and grade terms are combined as shown in these examples: weak fine granular; moderate medium columnar; strong coarse subangular blocky.

Consistence.--Soil consistence is the relative mutual attraction of the particles in the whole soil mass or their resistance to separation or deformation. Degrees of consistence are described in separate terms for three standard moisture states: dry, moist, and wet (Table 2.4).

1/ Australian soil scientists now use the term apedal to avoid the contradiction of having a term like structureless for a grade of soil structure (Butler, 1955).

Table 2.4

Soil consistence terms

A. Consistence when wet.--Moisture at field capacity or higher.

1. Stickiness or adhesion.--Measured by pressing between thumb and fingers.

Nonsticky.--No adhesion.

Slightly sticky.--Adheres, but comes off easily.

Sticky.--Adheres and stretches somewhat.

Very sticky.--Adheres strongly and stretches decidedly.

2. Plasticity or cohesion.--Measured by ability of material to form a wire between thumb and finger.

Nonplastic.--No wire formable.

Slightly plastic.--Wire formed but easily deformed.

Plastic.--Wire formed, requires moderate pressure to deform.

Very plastic.--Wire formed, requires much pressure to deform.

B. Consistence when moist.--Measured by resistance to rupture between thumb and finger.

Loose.--Noncoherent.

Very friable.--Crushes easily under very gentle pressure.

Friable.--Crushes easily under gentle pressure.

Firm.--Crushes under moderate pressure.

Very firm.--Barely crushable.

Extremely firm.--Cannot be crushed.

C. Consistence when dry.--Measured by resistance to rupture between thumb and fingers.

Loose.--Noncoherent.

Soft.--Breaks to powder or individual grains under very slight pressure.

Slightly hard.--Weakly resistant; easily broken.

Hard.--Moderately resistant; barely breakable.

Very hard.--Very resistant; not breakable.

D. Cementation.--Refers to brittle hard consistence caused by some cementing substance other than clay, such as calcium carbonate, silica, or oxides or salts of Fe, Mn, and Al.

Weakly cemented.--Brittle and hard but can be broken in hands.

Strongly cemented.--Brittle and hard; cannot be broken in hands but is easily broken by hammer blow.

Indurated.--Very strongly cemented; brittle and extremely hard; does not soften under prolonged wetting; breaks only after sharp hammer blow.

Clay films.--The clay that is translocated by water moving through a soil profile--i.e., eluviated from a surface horizon such as the A horizon and illuviated in a lower horizon such as the B horizon--is mainly colloidal clay, mostly less than 2 microns in particle size. The translocation is mostly along channels provided by root pores and the cleavage faces between peds. The clay particles are generally platy and after translocation tend to be oriented parallel with the surface on which they are deposited, forming laminated films, called clay skins, clay films, clay flows, or tonhaütchen. These films are one of the best criteria of a weathering profile. They can be seen readily with a 10x to 20x power hand lens. Distinguishing features are: color and texture different from the ped interior; commonly somewhat shiny, irregular surface, with channels and flow lines as if formed by running water; pores emerging from the lower side of a ped that commonly have irregular lips where the clay film protrudes; and tracings of molds of roots. In coarse granular material such as sandy soils, the translocated clay commonly forms coatings surrounding the grains and in places, bridges between grains.

Clay films are most readily identified where they are well developed in medium-textured materials. They are increasingly more difficult to recognize as the films become thinner, and as the matrix material becomes more clayey, particularly if the clay is of the swelling variety. Structureless or massive soils lack peds and thus cannot have clay skins on ped surfaces; any translocated clay in them occurs as coatings on individual sand grains, bridges between grains, and along occasional pores. The minimum amount of clay translocation that can be recognized consistently by trained soil morphologists is considered to be about 10 per cent loss of clay from the surface horizon (with corresponding illuviation in the B horizon) for medium-textured soils; coarse-textured soils require a considerably higher percentage of the total clay.

Soil reaction.--A rather wide range of chemical and physico-chemical characteristics can be used to describe soil horizons. Most of these are determinable only in the laboratory; only the property called "soil reaction" is customarily determined in the field. Soil reaction refers to the degree of acidity or alkalinity (pH) of the soil material. Precise measurement of pH requires use of special electro-chemical cells; generally the glass electrode is used, in reference with a calomel electrode. Such measurements are generally done in the laboratory, but can be made in the field if necessary. Approximate pH can be easily determined in the field by use of pH indicator dyes. Such field determinations are standard practice for soil scientists and also for many geologists (Richmond, 1962a; Scott, 1963, Morrison, in press, a and b), because they provide information on one of the most important chemical criteria for differentiation of soil horizons.

Colorimetric pH indicators give approximations to the correct pH value, if used with care. They should be used with the narrowest possible soil-water ratios, because the pH of some soils may vary with the proportion of water present, becoming higher with increasing dilution. The Soil Survey Manual points out that in addition to errors resulting from uncertainties as to soil-water ratio, colorimetric indicators are subject to errors because of masking by the original soil color, temperature, absorption of the dye on soil colloids, and soluble salts.

Colorimetric field pH determinations, being only approximate values, generally are expressed in the following terms, instead of their numerical values:

Soil reaction terms	pH values
Extremely acid	below 4.5
Very strongly acid	4.5 - 5.0
Strongly acid	5.1 - 5.5
Medium acid	5.6 - 6.0
Slightly acid	6.1 - 6.5
Neutral[1]	6.6 - 7.3
Mildly alkaline	7.4 - 7.8
Moderately alkaline	7.9 - 8.4
Strongly alkaline	8.5 - 9.0
Very strongly alkaline	9.1 - and higher

[1] Neutrality is pH 7.0, but soils in this range are called neutral.

Normal soils range in pH from about 3.5 to a little above 9.5.

The pH is a measure of the intensity of acidity or alkalinity, not the capacity or total amount. Soils high in clay or organic matter have higher capacity (greater reserves of either acidity or alkalinity) and are said to be well buffered; those low in clay or organic matter are correspondingly low in reserves of acidity or alkalinity. Kaolinitic clay has lower capacity than illitic or montmorillonitic clay.

Soil classification

Geologists concerned with using weathering profiles stratigraphically have found it advantageous to identify the general character of various profiles according to one of the broader categories of one of the existing schemes of classification used by soil scientists. This has seemed preferable to developing an independent classification, in spite of the fact that no existing classification by soil scientists is wholly satisfactory for use by geologists.

In the U. S., two very different schemes of soil classification are in use. One has developed slowly from Russian and European roots and resembles various soil classification systems used throughout the world. It has six categories[1/] (order, suborder, great soil group, family, series, and type), of which only the three highest categories generally need concern geologists (Baldwin, Kellogg, and Thorp, 1938; Thorp and Smith, 1949). The latest classification in these three categories is shown in table 2.5.

The three orders of the present classification are defined as follows:

Zonal soils are those produced under normal conditions from well-drained parent material (not of extreme texture or chemical composition) acted on by climate and biological forces for enough time to produce well-developed soil characteristics.

Intrazonal soils have more or less well-developed characteristics that reflect some local factor of relief, drainage, or parent material that overbalances the normal effect of climate and vegetation.

Azonal soils are those in which the parent material has remained just about as it was originally, almost unchanged by any forces.

1/ An uppermost category proposed by Marbut (1927, 1935), consisting of two classes, Pedalfer and Pedocal soils, has been abandoned by U. S. soil scientists, although it is still used by many geologists and will be referred to repeatedly in this dissertation. Pedocals are formed in semi-arid (locally sub-humid) to arid climates and are distinguished by accumulation of alkaline earth (Ca and/or Mg) carbonates in the lower part of, or even throughout the soil profile. Pedalfers are formed in relatively humid climates and are characterized by absence of accumulation of alkaline earth carbonates, and generally by accumulation of Fe and Al silicates and/or hydrolyzates.

Table 2.5. Soil Classification in the Higher Categories[2]

		Great soil groups
Zonal soils	1. Soils of the cold zone	Tundra soils.
	2. Light-colored soils of arid regions.	Desert soils.
		Red Desert soils.
		Sierozems.
		Brown soils.
		Reddish-Brown soils.
		Calcisols.[1]
		Grumisols.[1]
	3. Dark-colored soils of semi-arid, subhumid, and humid grasslands.	Chestnut soils.
		Reddish Chestnut soils.
		Chernozem soils.
		Prairie (Brunizem) soils.
		Reddish Prairie soils.
	4. Soils of the forest-grassland transition.	Degraded Chernozem.
		Noncalcic Brown soils.
	5. Light-colored podzolized soils of the timbered regions.	Podzol soils.
		Gray Wooded or Gray Podzolic soils.[1]
		Brown Podzolic soils.
		Gray-Brown Podzolic soils.[1]
		Sol Lessives.[1]
		Sols Bruns Acides.[1]
		Red-Yellow Podzolic soils.
	6. Lateritic soils of forested warm-temperate and tropical regions. (Latosols)	Reddish-Brown Lateritic soils.[1]
		Yellowish-Brown Lateritic soils.
		Laterite soils.[1]
Intrazonal soils	1. Halomorphic (saline and alkali) soils of imperfectly drained arid regions and littoral deposits.	Solonchak, or Saline soils.
		Solonetz soils.
		Soloth soils.
	2. Hydromorphic soils of marshes, swamps, seep areas, and flats	Humic Gley soils[1] (includes Wiesenboden).
		Alpine Meadow soils.
		Alpine Turf soils.[1]
		Ando soils.[1]
		Bog soils.
		Half-Bog soils.
		Low-Humic Gley soils.[1]
		Planosols.
		Ground-Water Podzol soils.
		Ground-Water Laterite soils.
	3. Calcimorphic soils	Brown Forest soils (Braunerde).
		Subartic Brown Forest soils.[1]
		Western Brown Forest soils.[1]
		Rendzina soils.
Azonal soils		Lithosols.
		Regosols (includes Dry Sands).
		Alluvial soils.

[1] New or recently modified great soil groups.

[2] Modified from Soil Survey Staff, 1960.

The definition of the zonal and intrazonal orders (as well as of many of the suborders and great-soil-groups) are mainly in genetic terms and therefore are not entirely satisfactory for a taxonomic classification based on observable soil characteristics. Zonal soils generally are moderately to strongly developed as weathering profiles (except some tundra soils), and intrazonal soils generally are less well developed--although Solonetz Soils, Planosols, and Brown Forest Soils commonly are as well developed as weathering profiles as are many soils in the zonal order.

The category in this classification that is most useful to geologists is the great soil group, because this divides soils into comparatively few main divisions (currently about 40 are recognized in the U. S.), on the basis of their dominant morphologic characteristics (albeit also on climatic-vegetational characteristics in some cases). Most geologist-soil-stratigraphers classify their soils by great soil groups, and this practice is followed in this dissertation. The great soil group classification is not completely satisfactory for use by geologists, because it places too much emphasis upon the upper part of the soil profile, including the A horizon. This is desirable from the agronomic standpoint, but undesirable from the geologic one, inasmuch as the upper part of the soil profile is the most liable to secondary modification, including erosion.

Another drawback for classification by great soil groups is that only brief, commonly vague definitions of these units have been published (Marbut, 1927, 1935, 1951: Kellogg, 1936; Baldwin, Kellogg, and Thorp, 1938). This has led to confusion among geologists trying to apply these definitions in classifying soils (as well as to important differences of opinion among soil scientists). Moreover, the great soil group classification has undergone progressive modification, with redefinition of some units and addition or deletion of others.

Recently the U. S. Soil Conservation Service (Soil Survey Staff, 1960) published a radically different system of soil classification that is scheduled to supersede the old classification. This organization has attempted to develop a natural taxonomic (descriptive) classification that maintains maximum objectivity, unincumbered either by inferences as to soil genesis or by previous soil-classification terminology. An almost wholly new nomenclature therefore has been developed, beginning with that for various soil horizons and extending through the five higher of the six categories of soil classification. The six categories, in decreasing rank are: order, suborder, great-group, subgroup, family, and soil series. Table 2.6 gives the 10 orders and their approximate equivalents under the old classification.

This new classification scheme, which has been nicknamed the "7th approximation" by soil scientists (scheduled soon to be superseded by an "8th approximation") achieves a new level of perfection of descriptive non-genetic classification for soils of all kinds, and it is now the official classification of the U. S. Soil Conservation Service. It has met with considerable resistance abroad, however, even in Canada (Leahey, 1963). It is not likely to be used widely by geologists concerned with describing and classifying soils for soil-stratigraphic work, because it places even more emphasis on the surface horizon of soils than does the older classification; it is quite complex and requires considerable training in soil science before it can be used competently; and in many cases laboratory data are needed in order properly to classify a soil.

Table 2.6. Soil orders in the "Seventh Approximation" and approximate equivalents in great soil groups [of revised classification of Baldwin et al. (1938) and Thorp and Smith (1949)] 1/

Present Order		Approximate Equivalents
1.	Entisols	Azonal soils, and some Low Humic Gley soils.
2.	Vertisols	Grumusols.
3.	Inceptisols	Ando, Sol Brun Acide, some Brown Forest, Low-Humic Gley, and Humic Gley soils.
4.	Aridisols	Desert, Reddish Desert, Sierozem, Solonchak, some Brown and Reddish Brown soils, and associated Solonetz.
5.	Mollisols	Chestnut, Chernozem, Brunizem (Prairie), Rendzinas, some Brown, Brown Forest, and associated Solonetz and Humic Gley soils.
6.	Spodosol	Podzols, Brown Podzolic soils, and Ground-Water Podzols.
7.	Alfisols	Gray-Brown Podzolic, Gray Wooded soils, Noncalcic Brown soils, Degraded Chernozem, and associated Planosols and some Half-Bog soils.
8.	Ultisols	Red-Yellow Podzolic soils, Reddish-Brown Lateritic soils of the U. S., and associated Planosols and Half-Bog soils.
9.	Oxisols	Laterite soils, Latosols.
10.	Histosols	Bog soils.

1/ From Soil Survey Staff, 1960.

CHAPTER III

STRATIGRAPHIC ATTRIBUTES OF GEOSOLS

How a geosol resembles, and differs from, a rock-stratigraphic unit (formation, member, etc.)

It has been pointed out above that geosols are stratigraphic entities distinct from deposits or geomorphic surfaces on which they formed. Like rock-stratigraphic units, they are recognized and defined by observed physical (and sometimes chemical) characteristics rather than by inferred age or history; certain diagnostic characteristics distinguish them from rock and other soil-stratigraphic units. They are conformable or disconformable bodies with arbitrarily chosen boundaries, interbedded with or overlying depositional sequences. Geosols are defined locally, but may be traced as far as their definitive features can be recognized. They may differ somewhat in age (generally not greatly) from place to place, like rock-stratigraphic units.

Geosols differ from rock-stratigraphic units in several significant ways: A geosol is not an original deposit; it is a zone of alteration that commonly transgresses earlier deposits. In this respect it resembles a zone of hydrothermal alteration or a metamorphic facies. A geosol develops in any earlier deposits that are exposed at the land surface while it is forming, and it alters them to make a new deposit and stratigraphic unit. Thus, although geosols are assigned to specific stratigraphic intervals within the rock-stratigraphic succession of an area, they are developed not only on rock-stratigraphic units immediately older than their assigned stratigraphic interval, but also on any earlier deposits that are exposed at the land surface while this geosol is forming. In fig. 5.2, for example, note how the Churchill

Soil is developed locally on the Eetza Formation, even though it belongs stratigraphically above the Wyemaha Formation.

A geosol also differs from most rock-stratigraphic units in being formed mainly from underlying deposits *in situ*, rather than of material transported from elsewhere. Furthermore, because a geosol is defined on its pedologic or weathering characteristics, it commonly lacks the lithologic homogeneity that is characteristic of many rock-stratigraphic units. A given geosol tends to reflect in its lithology that of the parent material from which it formed. Where a geosol has formed on parent materials that differ considerably in lithology, many details of lithology of the geosol will differ correspondingly, yet the geosol will maintain its distinctive pedologic characteristics.

How a geosol resembles, and differs from, an unconformity

Geosols commonly are associated with unconformities, and commonly they are useful criteria for recognizing and dating unconformities, but they are not unconformities themselves. The term unconformity refers merely to a surface--either an erosion surface buried beneath younger deposits, a surface between bedrock and overlying residual deposits, or a surface of non-deposition beneath younger deposits. A geosol, on the other hand, is the material product of weathering. It commonly is developed on an unconformity, extending downward from it, and has transformed the original parent material into a new deposit. It therefore has thickness, whereas an unconformity does not, being merely a surface. Because a geosol forms under conditions of slope stability, it represents a diastem in the sense of a gap in the depositional record. This diastem normally extends throughout a given area, but locally it may be represented by deposits in the low-lying places where deposition took place during development of the geosol. Geosols contribute important information on geologic history of an area, because they record an interval that is not generally recorded by deposits.

Types of stratigraphic occurrence of geosols

Buried and relict geosols

Geosols occur in two principal stratigraphic associations: either buried beneath younger deposits (buried geosols) or remaining at the land surface (relict geosols).

Buried geosols.--Buried occurrences of geosols generally are called buried soils (e.g., Thorp, 1949, Richmond and Frye, 1957; Richmond, 1962a; Scott, 1963; and Morrison, in press, a). They are seen only locally, generally only in good exposures and where the geosol has been sheltered from complete erosion prior to burial, inasmuch as they commonly are somewhat eroded before burial by the agencies that deposit the covering sediment. Burial favors preservation by inhibiting the various erosive, biologic and other agencies of the surficial environment; only a few feet of covering sediment essentially removes the geosol from the surficial environment. A buried geosol locally may undergo chemical or physical modification by subsurface agencies (see chapter 4). In most localities in the western U. S., however, buried Quaternary geosols are not appreciably modified by subsurface agencies, and very rarely so as to lose their distinctive characteristics.

Weakly developed geosols are found only in buried occurrences, never as relict ones, except in the case of the youngest geosols, such as the Toyeh and Midvale and younger geosols in the Lake Lahontan and Lake Bonneville areas (see chapter 5). This is the only way they are preserved unmodified by subsequent stronger pedogenesis (see fig. 3.1). Thus, in the older parts of Quaternary succession the geosols that are weakly or moderately developed are known only from buried occurrences. Fig. 5.2 shows an example of how buried geosols give a fuller record of separate pedogenic intervals, especially the weaker ones, than do relict soils, in locations favorable for the preservation of such geosols by burial.

Fig. 3.1. Fundamental relative development relations of relict and buried geosols. In A, the earlier geosol 1 is weakly developed and is preserved only in buried occurrences, because where it is relict it is masked by the stronger development of the later geosol 2. In B, however, geosol 1 is more strongly developed than geosol 2, and it is preserved in both buried and relict occurrences. The weak geosol 2 generally can be distinguished only where it developed on new deposits (rock unit B) separating it from geosol 1; where it developed on relict occurrences of geosol 1 it generally is masked by the stronger development of the latter (although in places it may be manifested as a composite profile). In such situations, geosol 1 preserves its distinctive characteristics in both buried and relict occurrences.

Soil scientists now restrict the term paleosol to buried soils (buried geosols)(Ruhe, 1956; Ruhe and Daniels, 1958; Soil Survey Staff, 1960, p. 30), but some geologists (Hunt and Sokoloff, 1950; Hunt, 1953b) have used this term for any ancient soil, whether buried or not.

Relict geosols.--Relict occurrences of geosols are those that have remained exposed at the land surface since they formed, without burial by younger deposits. They were called relict soils by Thorp (1949), because they are the remains of ancient soil profiles, albeit with their upper portions somewhat modified by subsequent surficial processes. Relict geosols obviously have been exposed to the entire sequence of changes in their surficial environment from the inception of soil-profile formation to the present. Some soil scientists (for example, Nikiforoff, 1949, 1955) consider that any soil that is exposed at the present land surface is in equilibrium with the presentday suficial environment and is essentially a "living" soil. Geologic soil-stratigraphic studies show, however, that this concept has very limited validity (see chapter 7). If relict occurrences of a given geosol are traced to and compared with buried occurrences of same geosol, in most cases only the uppermost part of the relict profile shows appreciable secondary modification by biologic or geologic activity (any local erosion excepted), and the lower part of the relict profile does not differ significantly from this portion of the profile of the buried geosol.

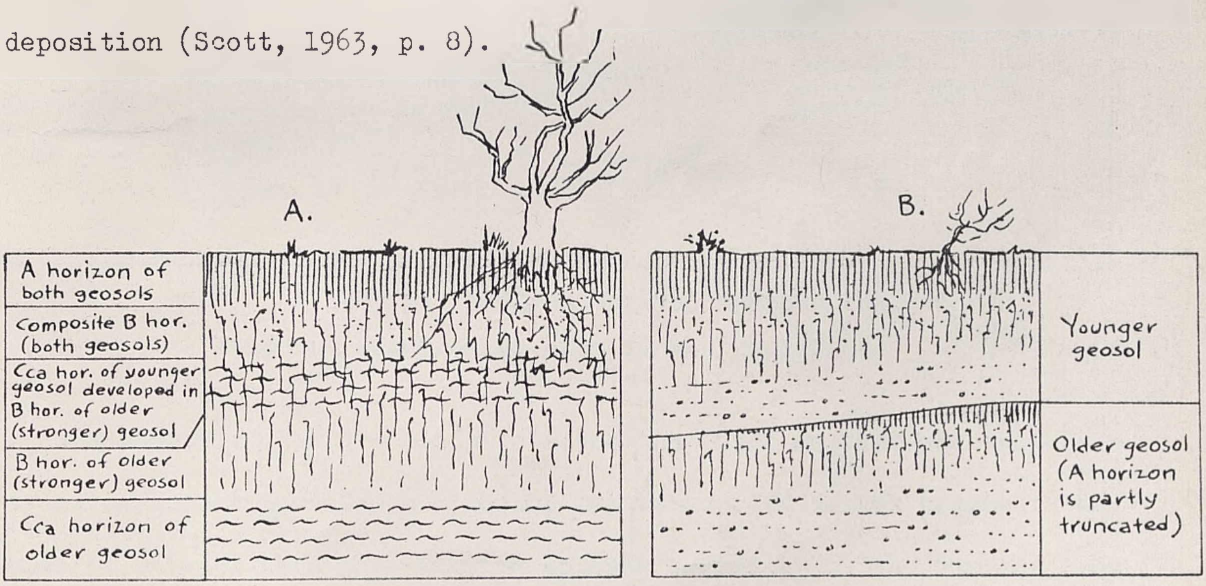
A strongly developed relict geosol is little affected by later weathering processes that produce only moderate or weak soil development on fresh parent material. This is because most of the soil material already is weathered past the products that would be developed by the weaker pedogenesis. Very strongly developed old relict geosols, such as those of Yarmouth age, have been subjected to several subsequent intervals of marked (as well as lesser) pedogenesis, yet typically these relict geosols preserve their identity wherever they have survived erosion. This principle also applies to relict occurrences of moderate and weakly developed young geosols, inasmuch as no intervals of stronger soil development have occurred since these geosols formed (see chapters 5 and 7). In such cases, however, the relict occurrences typically are somewhat more strongly developed than buried occurrences of the same geosol, because the climate of these lesser young soil-forming intervals was proportionately more similar to the climate of the intervening and subsequent times (e.g., see fig. 5.1).

Relict geosols are far more common than buried ones. In most environments they are widespread on all but the most modern geomorphic surfaces. They comprise the zonal and many of the intrazonal soils shown on soil maps, inasmuch as soils developed their chief genetic characteristics under ancient environments, not under modern ones. Because they are so ubiquitous, relict geosols are the chief tool of the soil stratigrapher. Once their precise stratigraphic position (age) and diagnostic features have been determined in an area, they can be used with as much confidence as buried geosols, provided that adequate safeguards are taken against confusing secondary modifications (see chapter 4). Relict geosols are particularly useful for differentiating and correlating various constructional and erosional landforms, for example, alluvial terraces and fans, lacustrine shore features, and glacial moraine and outwash features.

Superposed types of geosols, composite and compound

Some occurrences of geosols record more than one episode of pedogenesis; these will be called superposed occurrences of geosols. A relict geosol may have secondary modification not only of its A horizon but also of its B and even its Cca horizon because of subsequent pedogenesis. If the later pedogenic modification of the B and/or Cca horizons is sufficiently strong as to develop distinctive characteristics of its own, the occurrence is said to be composite (Bryan and Albritton, 1943) (fig. 3.2, A). Examples of such modification include chemical or physical changes in the B horizon of the original geosol (including deposition of calcium carbonate), leaching and/or redeposition of calcium carbonate in the B3ca and uppermost Cca horizons, etc. It should be noted that a composite occurrence is not considered to be a single geosol, but two different geosols that are partly merged together but still separately identifiable by means of their individual diagnostic characteristics. (In rare cases distinctive features are preserved that permit differentiation of more than two separate geosols in a composite profile.)

Composite geosols generally are identifiable only locally where strong or very strongly developed geosols have been exposed (relict) at the land surface during subsequent marked pedogenesis. Commonly a thin increment of new parent material has been added (e.g., as loess or slopewash) on top of the original geosol prior to the secondary pedogenesis, providing fresh material for weathering. As long as the profiles of different age overlap each other, they meet the definition of being composite. Composite geosols are common in some areas, rare in others. In the Kassler quadrangle, Colorado, almost every geosol shows secondary calcium carbonate deposition (Scott, 1963, p. 8).



A. Two composite geosols. (Illustrating a common type of occurrence in Pedocal areas).

B. Two compounded geosols.

Fig. 3.2
Superposed geosols.

Another type of superposition occurs where one geosol has formed closely above another, but separated by enough intervening sediment so that the two profiles do not overlap, or overlap very slightly. In such occurrences the lower profile, of course, is buried; nevertheless, if the profiles are nearly contiguous, separated by only a few inches to several feet of intervening not-soil deposits, it may be convenient to call the ensemble a compound or compounded occurrence. An occurrence where two or more <u>different</u> geosols are superposed (fig. 3.2, B) is said to be <u>compounded</u>; an occurrence where separate weathering profiles are superposed that are all related to a single geosol (fig. 3.3) is said to be a <u>compound</u> occurrence of this geosol.

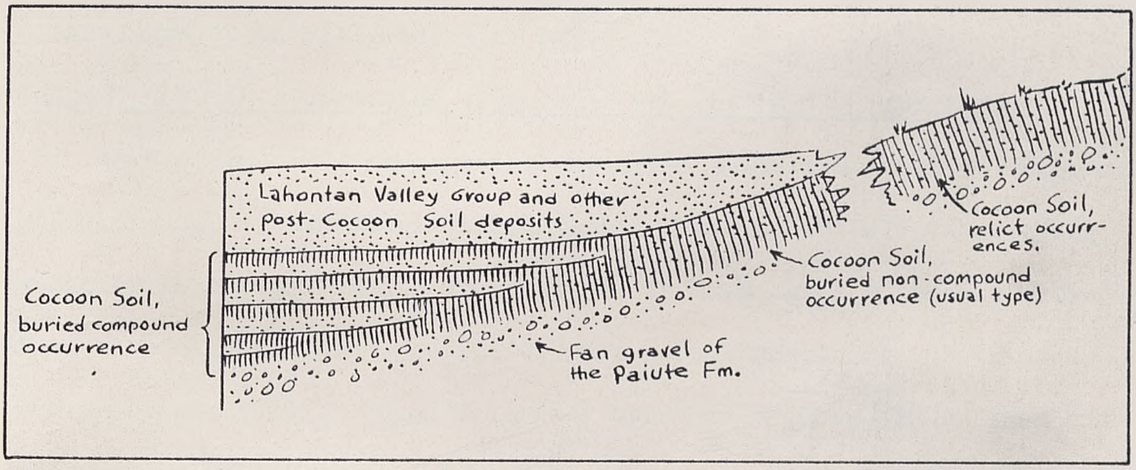


Fig. 3.3. Diagram showing compound and non-compound occurrences of a single geosol.

Any number of geosols can be stacked one above another and the whole succession termed a compound(ed) geosol occurrence, if enough sediment was deposited between each soil-forming episode to result in the various profiles being distinctly though not excessively separated. Commonly some or all of the individual weathering profiles in a compound(ed) occurrence are truncated; i.e., their upper parts are somewhat eroded. Also, some of the profiles may overlap one another sufficiently to qualify as composite occurrences. In many cases it is possible to trace a composite occurrence laterally into a compound(ed) occurrence of geosols, affording opportunity to identify the individual geosols.

Compound(ed) occurrences of geosols are most likely to be found in situations where minor increments of sediment tend to be deposited between the soil-forming intervals. These generally are low-lying locations sheltered from strong erosion or deposition, such as the lower parts of colluvial slopes and alluvial fans, outer margins of flood plains, swales in hilly terrain, etc.

General note on designation of geosols vs. weathering profiles

In all types of stratigraphic occurrence--buried, relict, composite and compound(ed)--the soil-stratigraphic units do not always meet the stratigraphic requirements of a geosol; in such cases a less specific term such as weathering profile or para-geosol should be substituted. The appropriate usage depends largely upon how a geosol is defined and its normal mode of stratigraphic occurrence (or, if an informal soil-stratigraphic unit, how it normally occurs and is used). Fig. 3.3 shows a typical example. The Cocoon Soil (Geosol) is a formal soil-stratigraphic unit that normally occurs as only a single weathering profile in both relict and buried occurrences. In a few low-lying localities (generally former stream floodplains), however, this soil-stratigraphic unit is represented by several (locally as many as 12 separate weathering profiles, that can be said to comprise a compound occurrence. Inasmuch as the whole compound occurrence is equivalent to the unit defined as the Cocoon Soil, the individual weathering profiles cannot themselves be called geosols; they can only be designated as weathering profiles within (and part of) the single Cocoon Soil.

Another example: The weathering profiles developed during the three main pre-Wisconsin interglaciations (the Afton, Yarmouth, and Sangamon Interglaciations) commonly are so similar in their degree of development that they can be identified as individual geosols only where they occur buried in stratigraphic succession in the same exposure. In relict occurrences they commonly look so much alike that it cannot be ascertained which of the three weathering profiles is represented, as well as the number of them that may be superposed on each other (fig. 4.6). In cases where their specific age and stratigraphic entity cannot be identified, probably they cannot properly be termed geosols, but should be designated merely as weathering profiles.

The periodic nature of weathering intervals; designation of geosols in terms of erosion-weathering cycles

A valuable concept for understanding (and designating) the stratigraphic-age relations of geosols is the "periodic phenomena in landscapes" concept advanced by Butler (1958a, 1958b, 1959). This concept now has come to be widely accepted by Australian soil scientists as an important advance in the study of soils (Van Dijck, 1958, 1959; Jessup, 1960a, 1960b, 1961;

Langford-Smith, 1962; Walker, 1962a, 1962b; Butler, 1960, 1961). In my opinion it deserves still wider acceptance; however, I propose several modifications to Butler's original terminology in order to make it more acceptable to U. S. stratigraphers and to my own ideas of the requirements of soil stratigraphy.

Butler has demonstrated that intervals of soil development have been discontinuous; i.e., they have recurred periodically and have alternated with intervals of active erosion and deposition. Therefore, the rates of erosion and deposition and of soil development have differed considerably with time. This viewpoint is diametrically opposed to Nikiforoff's (1949, 1955), who advocates that a steady state of dynamic equilibrium is maintained in soils, with soil development keeping pace with changes in erosion and deposition whenever "maturity" of the soil is attained. (Nikiforoff's concept is further discussed in chapter 7.)

In order that a weathering profile can develop, erosional losses and depositional additions to the site must be essentially nil for a sufficiently long time--in other words, erosional-depositional stability must be maintained for a certain interval (Frye, 1951; Hunt, 1953b, p. 42-43; Butler, 1959). Occurrence of buried soils demonstrates that the stability intervals, when the soils formed, were followed by intervals of active deposition (due to erosion elsewhere) when deposition was so rapid that the time of exposure of each depositional increment was too short for any soil development to take place on it.

The principle of periodicity recognizes recurrent cycles, beginning with an interval of instability with active erosion and deposition, followed by an interval of stability, with weathering profile development. These cycles can be considered to be time-stratigraphic units. They are here called "erosion-weathering cycles." (Butler calls them "soil cycles", but this term seems inappropriate because it emphasizes only one of the dual aspects of the cycle.) Each full cycle consists of an instability interval that is recorded at least locally by a deposit and(or) an erosional surface (unconformity); this interval is followed by a stability interval that is recorded by a stability surface and a weathering profile. In other words, a cycle commences with erosion or deposition which creates either a new erosion surface (unconformity) or buries the former stability surface, and which is followed by maintenance of a new stability surface and development of a weathering profile beneath this surface.

Butler proposes that these cycles of erosion-deposition to stability-weathering logically are time units. He called the time unit represented by a single cycle a K cycle (after the Greek Khronus = time), and he defines a K cycle as "an interval of time covering the formation, by erosion and(or) deposition, of a new landscape surface, the period of development of soils on that surface, and ending with the renewal of erosion of and(or) deposition on that surface". I doubt, however, whether K cycles are valid geologic time units, inasmuch as they probably are generally (though not necessarily invariably) controlled by climatic changes, and these changes and their erosional-depositional manifestations may vary considerably from one climatic terrain to another, with altitude and latitude, etc. In my opinion, K cycles qualify as geologic-climatic units (under the new Stratigraphic Code), rather than as geologic-time units. Because they are bounded by weathering profiles, however, they may also qualify as time-stratigraphic units, in view of the fact that weathering profiles (geosols) are para-isochronous (see chapter 6).

Butler's K-cycle classification of weathering profiles, surficial deposits, and erosion surfaces is adopted in this dissertation, with certain modifications. (Whether the name "K cycle" is retained is immaterial; the important contribution is the basic concept of periodically recurring erosional-weathering cycles.) Following Butler, the individual K cycles are denoted by numerical subscripts, starting with K_0 for the present and going backward in time. The unstable phase is indicated by subscript u; the stable phase by subscript s, thus: K_{1s}, K_{3u}(fig. 3.5).

Butler mentions the possibility of only one instability deposit or erosion surface during a single K cycle. In some areas, however, several kinds of erosional-depositional processes may operate successively during a single general instability interval. Evidence that this commonly was so during glacial-interglacial (and pluvial-interpluvial) cycles is recorded in many parts of the western U. S., notably in the glaciated and periglacial areas (fig. 3.4). Thus, in some areas more than one instability deposit and even more than one erosion surface may occur within the same K cycle. Such multiple units within the same main instability interval are designated by subscript lower-case letters, a, b, c, etc., from oldest to youngest.

It is well to remember that not all K cycles go to completion: erosion-deposition commonly does not destroy all the former stability surface and soil; likewise, the manifestation of a stability-weathering interval may be weak.

As a time-stratigraphic unit, the K cycle is of low rank, commonly much shorter than stage-age. Its duration is not a criterion in its determination. K cycles are superimposed upon geomorphic erosion cycles, and the stage of completeness of a geomorphic cycle ("stage" in geomorphic terminology)

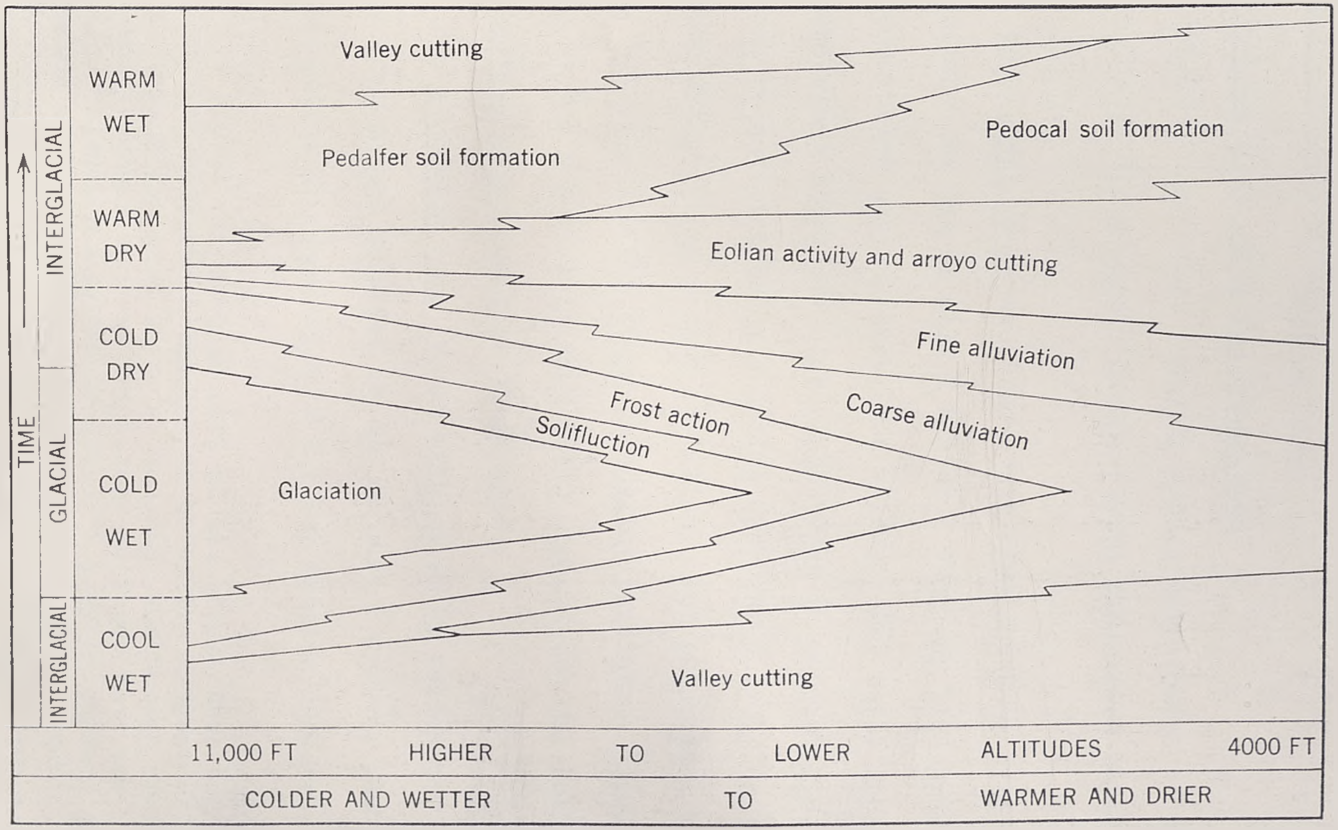


Fig. 3.4. Inferred changes in climate and in dominant surficial processes with time during a typical glacial-interglacial cycle in the La Sal Mountains, Utah (Richmond, 1962a). (More recent chronologic data indicate that the typical soil-forming interval lasted only about 1/10 of the total cycle, instead of about 1/5 as shown.)

has no relation to occurrence of K cycles.[1/] Tricart (1956) and Cotton (1958) have used the term "morphogenetic system" for comprehensive systems of Pleistocene landscape activity similar to Butler's K cycles, but they did not subdivide and classify these systems.

K-cycle zones in surficial landscapes

Periodic landscape activity causes development of characteristic zones in surficial landscapes. Butler (1959) classifies these zones as follows: (Fig. 3.5):

Persistent zones are those that have not been affected by erosion or deposition for one or more K cycles. They may lie topographically either above or below active zones (see below). The soils and stability surfaces in persistent zones are older than those in adjoining active zones; their age can be ascertained by counting back the K-cycle units from the youngest K-cycle unit. (This method is reliable for persistent zones beyond deposition, but may give too young an age for persistent zones beyond erosion, owing to the possibility that later depositions may bury earlier ones.)

Active zones comprise three types:

The sloughing zone (which might also be called the erosion zone) is where erosion has occurred in one or more K cycles, that has caused removal of its debris and formation of a fresh erosion surface. The erosion can be

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[1/] Butler (1959) proposed as part of his scheme that the specific kind of surface that concerns a pedologist be called a "groundsurface". A single groundsurface includes the new deposits, weathering profile (soil), and erosional-depositional and stability surfaces that are related to a single K cycle. Although I advocate Butler's K-cycle concept and many of its appurtenant features, I consider that the term "groundsurface" is unfortunate. This term is not likely to be adopted by stratigraphers working within the concepts of the revised Stratigraphic Code (Amer. Comm. on Stratigraphic Nomenclature, 1961). For example, a "groundsurface" is not really a surface; it has thickness, inasmuch as it includes a weathering profile and, commonly, underlying deposits resulting from an instability interval. Thus, a groundsurface mixes geomorphic, rock-stratigraphic, and soil-stratigraphic terminology.

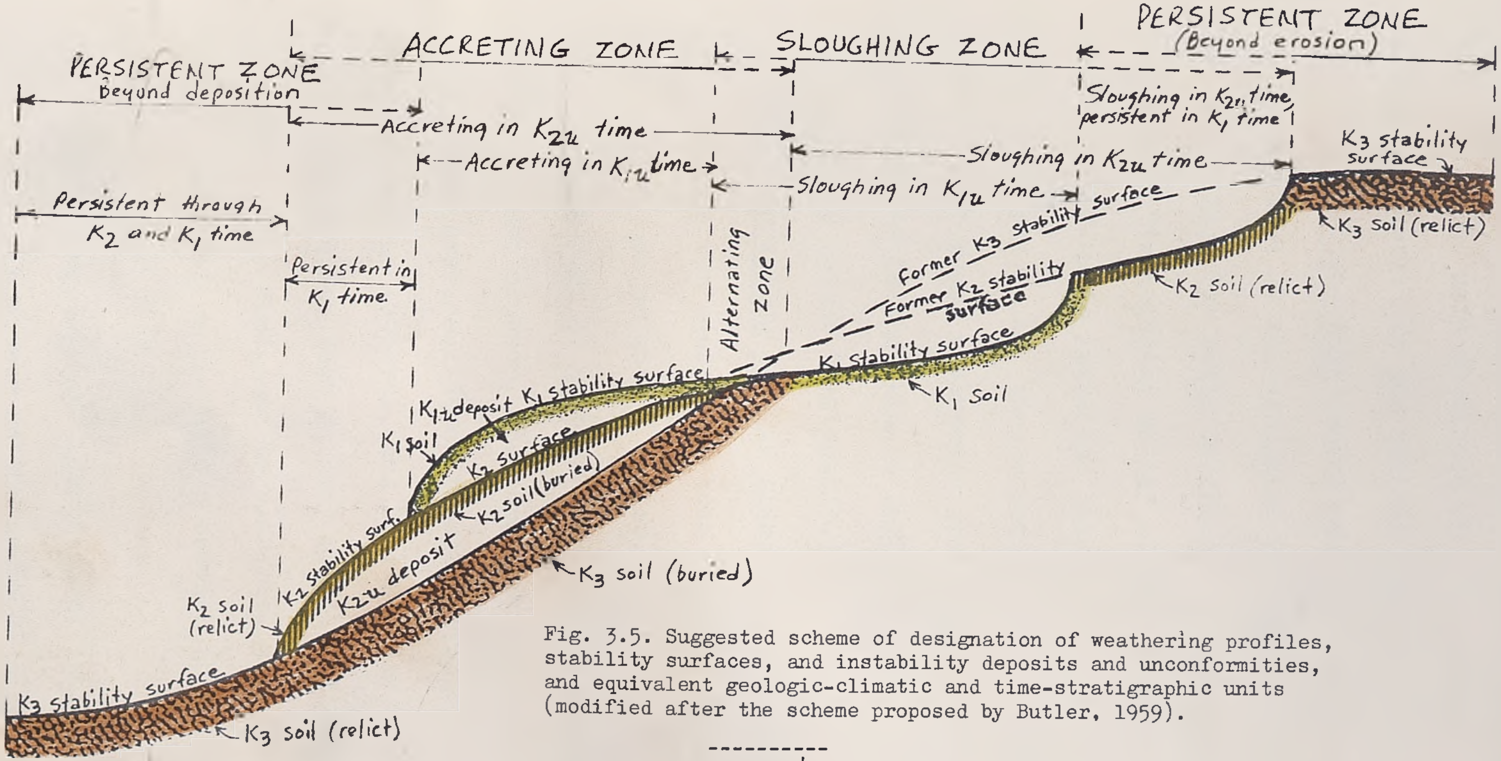


Fig. 3.5. Suggested scheme of designation of weathering profiles, stability surfaces, and instability deposits and unconformities, and equivalent geologic-climatic and time-stratigraphic units (modified after the scheme proposed by Butler, 1959).

Geologic-climatic (and time-stratigraphic) units		Rock-stratigraphic, soil-stratigraphic, and geomorphic surface units
K_1 erosion-weathering cycle (youngest)	K_{1s} stability interval or K_1 weathering optimum)	K_1 weathering profile (soil) and K_1 stability surface
	K_{1u} instability interval	K_1 deposit and K_1 erosional surface (or K_1 unconformity)
K_2 erosion-weathering cycle	K_{2s} stability interval (or K_2 weathering optimum)	K_2 weathering profile (soil) and K_2 stability surface
	K_{2u} instability interval	K_2 deposit and K_2 erosional surface (or K_2 unconformity).
K_3 erosion-weathering cycle (oldest)	K_{3s} stability interval (or K_3 weathering optimum)	K_3 weathering profile (soil) and K_3 stability surface
	K_{3u} instability interval	K_3 deposit and K_3 erosional surface (or K_3 unconformity)

Time

by any of the surficial types: mass-wasting (solifluction, landsliding, congeliturbation, etc.), hillslope washing, stream degradation, eolian deflation, etc. The soil developed on this erosion surface during the ensuing stability interval is the typical sedentary soil of the literature.

The <u>accreting zone</u> (which can also be termed the <u>deposition zone</u>) is the counterpart of the sloughing zone and has received deposition in one or more K cycles. The soils developed on the various depositional layers are the typical "transported soils" of the literature. Where they are buried by multiple K-cycle activity, they are either compound(ed) (buried soils of the literature), if the accreted layer is thick enough to fully separate the individual weathering profiles, or they are composite, if the accreted layer is too thin to fully separate the weathering profiles. (Butler calls the latter "partsoils".)

The <u>alternating zone</u> is transitional between the sloughing and accreting zones, and is the zone where both erosion and deposition have occurred at the same site in one or more K cycles.

Although Fig. 3. 5 shows these various zones schematically in terms of hillslope slump or wash, the basic concepts and terminology apply to other lithogenetic erosional-depositional terrains, such as glacial, fluvial, lacustrine, marine littoral, and eolian. Any of the various zones may range in width from a few feet to many miles. It also is important to note that agencies such as wind, and to a degree, also ice, act independently of level, so that the relative elevations between zones shown in fig. 3. 5 need not apply.

Various K cycles have identities of their own. They can be differentiated and distinguished on the basis of different characteristics of their components, such as differences in lithology of deposits, in character of weathering profiles, lateral extent and degree of unconformities due to variation in type and intensity of operation of the agencies of erosion, transport, and deposition, and weathering.

Maignien (1960), on the basis of his observations in Europe and tropical Africa, pointed out that the periodic nature of erosion-weathering cycles is most clearly manifested in the stratigraphic record preserved in semiarid and arid regions and at temperate latitudes. It is least clear in the humid tropics, partly because here the amplitude of Quaternary climatic changes (and resulting erosional-depositional changes) was much less than in temperate latitudes, and also because pedogenic changes tend to progress so rapidly in the tropics that they tend to obliterate the earlier records.

In conclusion, the K-cycle scheme offered by Butler promises to be a useful framework for considering geosols in their stratigraphic and geomorphic relations--in other words, for soil-stratigraphic work. The proper procedure for study is first to differentiate the succession of whole K cycles, then to differentiate and study the component parts of the various K cycles, including the individual geosols.

Facies variants of geosols

Soil facies is an important concept in soil stratigraphy that differs from usage by soil scientists. A given geosol commonly ranges considerably in its characteristics on account of changes in various factors which affected its formation. The chief soil-forming factors that can vary in different parts of an area and consequently cause variation in the original characteristics of a geosol are: climate (particularly temperature and precipitation, as with altitude in a mountain range), vegetation and other organisms, relief (slope, etc.), drainage, and parent material (especially its texture and composition).

Such soil changes have recently been termed soil facies by several geologists (Richmond, 1962a, in press; Scott, 1963; Morrison, 1961c, and in press, a and b), by analogy with the use of this term for lateral changes in rocks. These geologists use standard great soil group names (such as Brown Soil, Desert Soil, Chestnut Soil, Chernozem, etc.; see chapter 2) to describe individual soil facies, and this practice is followed in this dissertation.

Climate is the principal determinant of the broad pattern of changes in soil characteristics that is the basis for classification into great soil groups (the vegetation factor is also, of course, closely dependent upon climate). Temperature is influenced mainly by latitude and altitude. Mean annual temperature changes about 27° F between the Mexican and Canadian borders of the U. S. Likewise, mean annual temperature decreases about 3.3° F for each 1,000 feet rise in elevation. Precipitation patterns are less regular, being influenced not only by large-scale climatic patterns in the general atmospheric circulation, but also by orientation and configuration of mountain masses and other topographic features. The geographic distribution of great soil groups essentially follows that of regional climatic provinces (figs. 3.6, 3.7, and 3.8). In regions of low relief, climatic changes are mainly those due to latitude and general atmospheric circulation, resulting in extensive climatic provinces and also in extensive areas of individual great soil groups. In mountainous areas, however, climatic provinces are telescoped into narrow altitude zones, resulting in marked changes in great soil groups (soil facies). Thorp (1931) described a succession of soils concentrically ringing the Big Horn Mountains of Wyoming, with pedocals (calcium carbonate-enriched soils) near the base and podzols near the top (fig. 3.9). Richmond (1962a) described a succession in the La Sal Mountains area, Utah, that is typical for the middle Rocky Mountain region (fig. 4.8): from Sierozems in the semiarid lowlands near the Colorado River to Brown, Brown Forest, and Brown Podzolic at successively higher altitudes. Marked changes in soil facies require that certain precautions be taken if geosols are used for local rock-stratigraphic differentiation and correlation (see chapter 4).

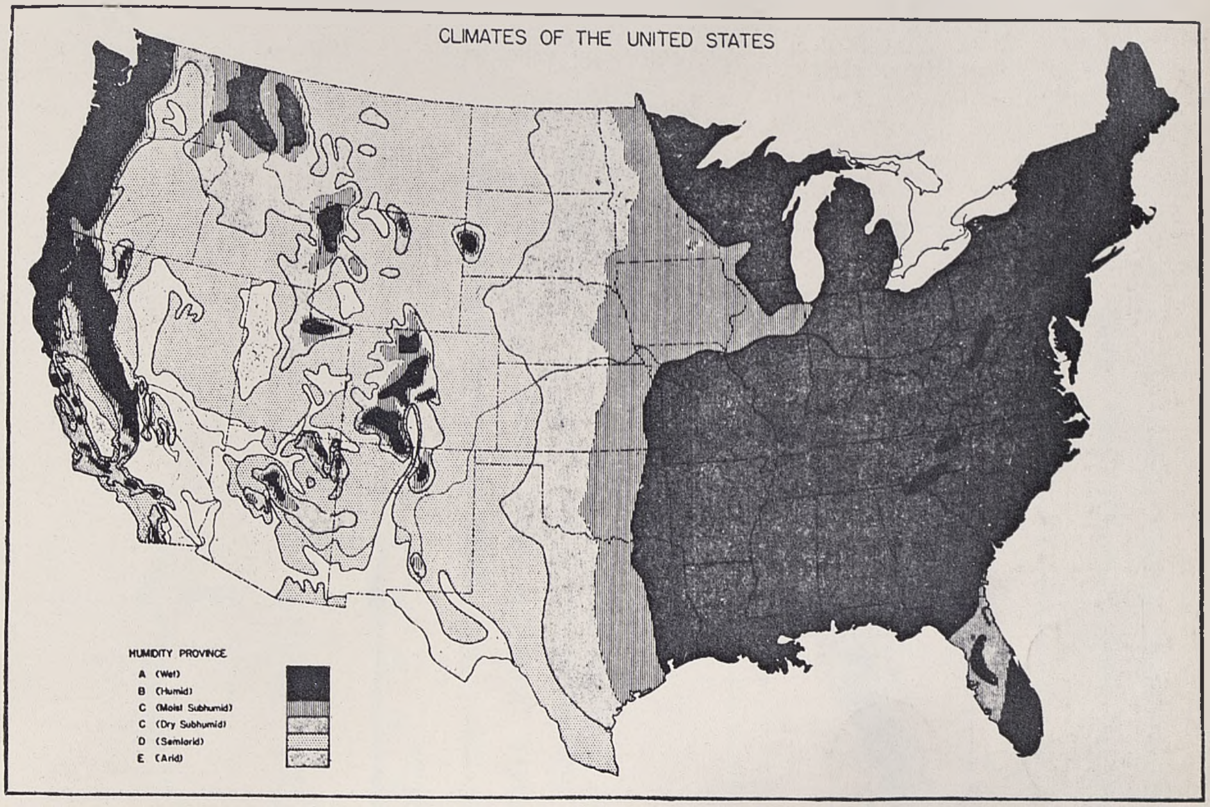


Fig. 3.6. Distribution of moisture regions in the U. S. according to Thornthwaite (after Jenny, 1941).

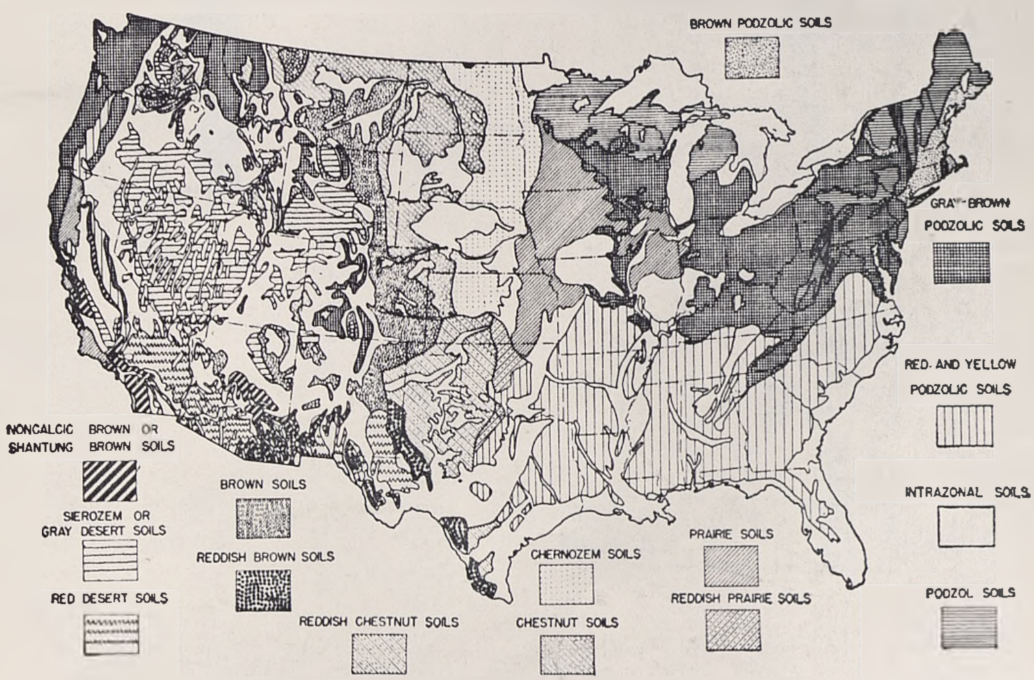


Fig. 3.7. Distribution of principal zonal great soil groups in the U. S. (after Jenny, 1941, after Kellogg, 1938).

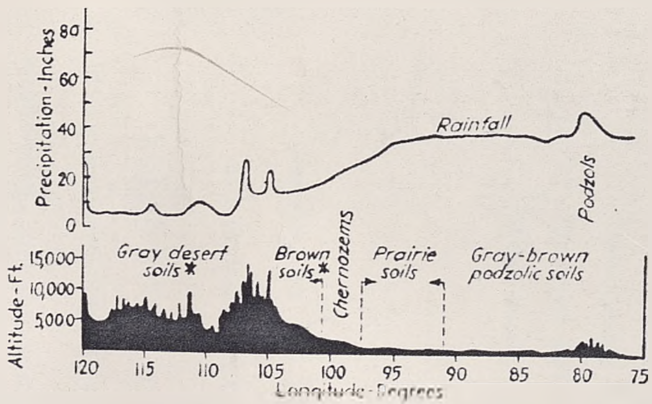


Fig. 3.8. Relief, precipitation, and zonal great soil group relations approximately along the 11°C. isotherm in the United States (modified after Jenny, 1941). Pedocals (mainly Desert and Brown soils) occur only below 4,500 to 9,000 ft altitude, depeding on local factors; pedalfers (mainly Brown Podzolic soils) occur at higher altitudes.

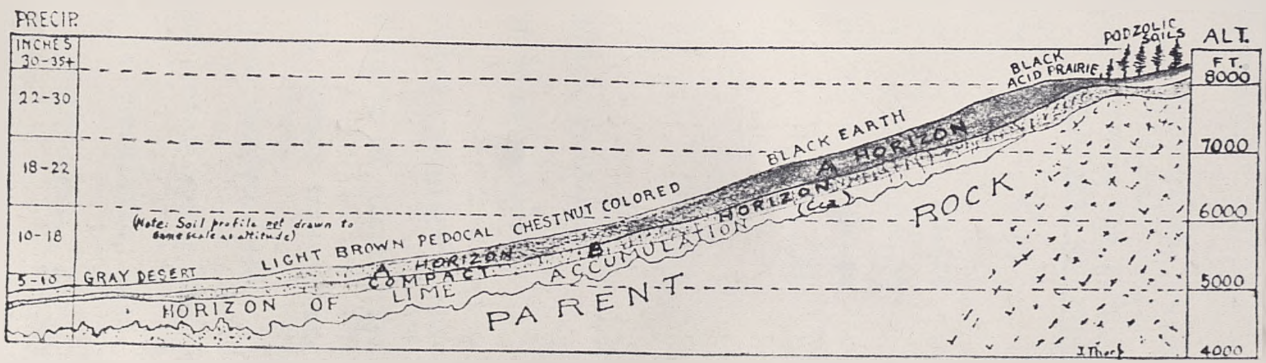


Fig. 3.9. Soil-facies changes from desert to humid mountain top. Western slope of the Big Horn Mountains, Wyoming (after Thorp, 1931).

Marbut (1935) gave the precipitation limits for various main great soil groups in the temperate U. S. as follows:

Table 3.1

Marbut's precipitation limits for various main zonal great soil groups

Zonal great soil groups	Approximate annual rainfall limits along the 11°C. isotherm, in inches
Gray Desert soils	<15
Brown soils (arid Brown and Chestnut soils)	15--20
Chernozems	20--30
Prairie soils	30--40
Gray-Brown-Podzolic soils . . .	30--50

Summary of the principal stratigraphic attributes and modes of stratigraphic occurrence of geosols

(1) A geosol is the fundamental soil-stratigraphic unit. It is a laterally traceable and mappable layer of distinctly weathered, predominantly mineral material, formed immediately beneath and generally parallel with the land surface, that maintains a consistent stratigraphic (age) relationship to older and younger deposits with which it is associated, and is defined and used on the basis of this relationship. It must meet at least the minimum requirements for degree of

weathering and stratigraphic entity. A geosol is given a local geographic name that distinguishes it from other stratigraphic units and from pedologic units such as soil-series units.

A unit that evinces distinct surficial weathering but does not meet the strict stratigraphic requirements of a geosol may be called a para-geosol or a weathering profile (or, less desirably, a soil). A para-geosol has all the stratigraphic attributes and modes of stratigraphic occurrence that are described for geosols, and is itself a soil-stratigraphic unit.

(2) A geosol is a distinct stratigraphic entity; it is distinct from deposits, depositional surfaces, or erosional surfaces (unconformities) on which it formed. It has certain attributes like, and others unlike, those of both rock-stratigraphic units (formation, member, bed) and unconformities.

(3) A geosol may occur either buried by younger deposits (<u>buried</u> geosol) or exposed continuously at the land surface since it formed (<u>relict</u> geosol). If a weak geosol is exposed (relict) at the land surface during a subsequent interval of stronger weathering, it loses its identity because it becomes transformed into (and masked by) the stronger weathering profile developed by the later pedogenesis. Thus only the stronger old geosols occur as relict ones; the weaker old geosols may, however, be preserved where burial has protected them from masking by subsequent stronger pedogenesis.

(4) Two or more geosols (weathering profiles) may be superposed in two ways, either with their weathering profiles overlapping (with distinct pedogenic modification of the older profile by the younger), termed a <u>composite</u> occurrence, or with their profiles slightly separated by intervening deposits, termed a <u>compound</u>(<u>ed</u>) occurrence.

(5) A geosol forms at the earth's surface when processes of weathering exceed those of deposition or erosion. It records a time of general slope

stability, although minor deposition or erosion may go on locally. The weathering processes that develop a geosol involve the interaction of six principal factors: climate, time, living organisms, relief, drainage, and parent material. The environmental factors, particularly climate, vegetation, drainage, and parent material, produce lateral changes in the type of weathering profile that is developed, in different parts of an area. These lateral changes in a geosol are called soil facies.

(6) Geosols commonly differ from one another in degree of development (p. 30). Each geosol, however, maintains the same general degree of development relative to the other geosols wherever it occurs. This holds true in spite of changes in soil facies (p. 64). The more strongly developed geosols are among the best stratigraphic markers.

(7) In every area where the age relations of sequences of Quaternary geosols have been determined by comprehensive stratigraphic study, invariably each geosol is of distinct geologic age, and the interval when it formed was short relative to intervening non-soil-forming intervals. This concept of the age relations of geosols differs importantly from the concept held by many soil scientists--that the age of a soil is the total time since the cessation of deposition of the parent materials on which the soil formed, or the total time that the land surface on which the soil formed has been exposed subaerially.

A geosol invariably is an ancient soil, not a modern soil, at least at temperate latitudes. This assertion can be made because reliable stratigraphic studies have shown that the amount of pedogenic weathering that has occurred in the last few centuries has been too slight to have developed the attributes of a geosol (see chapter 7).

CHAPTER IV

USE OF GEOSOLS AS STRATIGRAPHIC MARKERS WITHIN LIMITED AREAS FOR LOCAL DIFFERENTIATION AND MAPPING OF QUATERNARY DEPOSITIONAL UNITS AND LANDFORMS

How to identify and differentiate geosols in the field

Most geologists have little difficulty in recognizing a strongly developed geosol even if they have had little experience in soil stratigraphy, but commonly they find the weaker weathering profiles difficult to identify reliably. Following are a few characteristics that are particularly diagnostic of profiles of weathering. Full understanding and effective use of these criteria develops slowly and only after considerable practice in identifying and describing weathering profiles in the field.

Criteria of possible geosols in "weathered" exposures

First will be listed a few general criteria that have been found to be useful for "spotting" possible geosols, especially buried ones, even in "weathered" exposures and at moderate distances (table 4.1). They are intended to be used only as preliminary field guides, to facilitate rapid field reconnaissance. They must be supplemented by closer observation in the light of the more specific criteria for identification and comparison of geosols, which are discussed in the next section.

Guide for detailed field observation of possible geosols

Closer inspection and evaluation of possible geosols should be made in clean exposures of the fullest profiles available. Generally the older geosols are more or less eroded or otherwise secondarily modified (see last part of this chapter), and considerable search is needed to find comparative-full, well preserved profiles. In some situations the geologist may be able to identify a given geosol if only the lower part of its profile is preserved,

Table 4.1. General guides to possible geosols (especially buried ones), for use in rapid field reconnaissance

General guide feature	Remarks	Similar "not-soil" features that sometimes are confused with weathering profiles
1) Possible A horizon: Dark gray to black organic horizon.	1) Generally conspicuous where present, but commonly eroded, leached, and(or) oxidized (and bleached) in buried occurrences.	1) Organic horizons do not themselves indicate a weathering profile. They may overlie either a weathered B horizon or just unweathered parent material (A-C profile).
2) Possible B horizon: a) Color: various shades of gray-brown, yellow-brown, brown, reddish-brown and red.	2a) One of the most useful general guides, but may be masked or modified by strongly colored parent material, or by excessive organic matter and (or) gleying because of poor drainage.	2a) Similar color zones may be caused by deposition of iron oxides by ground water, or by oxidation by heating by fire or by lava flows.
b) Evidence of clay concentration, as by "checked" appearance of "weathered" exposures.	2b) Generally evident only in medium-& coarse-textured parent materials that have been strongly weathered. Most noticeable in montmorillonitic B horizons. May be masked by clayey parent material.	2b) Unweathered clay commonly develops similar checking on exposure, particularly if montmorillonitic.
c) Closely jointed appearance suggestive of blocky or prismatic soil structure.	2c) Generally evident only in the more strongly developed weathering profiles.	2c) Some unweathered clays develop similar close jointing on exposure and desiccation.
3) Possible Cca horizon (or duripan): a) Color: White, pale gray, pale brown, or pale pinkish.	3a,b) The more strongly developed Cca horizons commonly are very conspicuous. Their whiteness generally reflects the degree of calcium carbonate (or silica) concentration. Efflorescent surficial concentration of calcium carbonate ~~(see next section)~~ may accentuate the conspicuousness of weak or moderately developed Cca horizons, in arid areas.	3a) Whitish zones of efflorescent salines, viewed from a distance, may be mistaken for Cca horizons.
b) Cementation: Cap-rock effects and other manifestations of resistance to erosion.		3b) Zones of groundwater $CaCO_3$ deposition or of lacustrine or spring tufa deposition ~~(see next section)~~ may closely resemble pedogenic Cca horizons.

but caution is needed. Such identification is least reliable with pedalfer soil facies, but may be feasible for strongly developed pedocals, because their Cca (caliche) horizons not only tend to resist erosion but also commonly possess diagnostic characteristics of their own.

In any case, observation should be made in fresh exposures. Geologists unaccustomed to working with soils may not realize how important it is to observe them in their pristine state--not in weathered, slumped exposures. The mark of a soil-oriented geologist is his digging tools--shovel, pick, and if necessary, soil auger--for he realizes that soils can be properly observed only in fresh exposures, developed by fully cleaning off available natural or artificial cuts or by digging pits through the whole soil profile if no exposures are available.

A horizon.--The A horizon is of little or no utility for soil-stratigraphic correlation. It may or may not be part of a weathering profile. In relict occurrences it is especially prone to erosion and(or) secondary modification by organisms, etc. In buried old soils this horizon generally is lacking. Commonly this is attributed to erosion of the A horizon; however, Scott (1963, p. 8) suggests that the absence of an organic layer may be due to leaching or oxidation of the organic matter, or to conversion to fixed carbon of the original organic matter. He observed in soils of the Kassler Quadrangle, Colo., a marked progressive decrease in grayness of buried A horizons with age, which could be due to either cause.

B horizon.--The B horizon generally is the most diagnostic and reliable horizon for identification of geosols. It is comparatively little affected by secondary modification such as by addition of humus or by superposition of later weaker soils. The main properties that are diagnostic are: color, texture, structure, consistence, clay skins, reaction, and thickness.

Color commonly is one of the most reliable criteria, particularly where parent materials tend to be reasonably uniform and neutral in color. In the Kassler Quadrangle, Colo. (where the parent materials generally meet this requirement, for example, Scott (1963, p. 6-7) found that the older very strongly developed weathering profiles, tentatively correlated with the Aftonian, Yarmouth, and Sangamon Interglaciations, have B horizons that are dusky red and reddish-brown. The B horizons of successively younger weathering profiles are progressively reddish-brown, yellowish-brown, and grayish-brown. Soil color obviously is less diagnostic where it is masked or modified by highly colored parent material. Furthermore, profiles that develop in poorly drained locations with intermittently high water table typically have so-called "gleyed" characteristics because of the reducing environment, and their B horizons are shades of gray, in places mottled with yellow or brown. In minimal weathering profiles color may be especially diagnostic of the B horizon where this horizon lacks discernible texture, structure, or consistence change; development of yellowish or brownish color below the A horizon due to slight oxidation of their parent material (the so-called "color B" horizon) may be the dominant characteristic of such weak profiles.

Secondary color changes that sometimes resemble those due to pedogenesis may be caused by deposition of iron oxides (limonite, goethite, hematite, etc.) by groundwater as coatings or cemented zones. Commonly, though not invariably, the ferruginous zones are at or near the base or top of a permeable bed or zone that is underlain or overlain by less permeable material. Such ferruginous zones may or may not crosscut the stratification, and they may or may not be obviously unrelated to a former land surface. Close inspection generally reveals that they lack the characteristics of weathered zones such colloidal clay coatings, skins and bridges, etc.

Rarely, heating will produce a secondary color change, as by oxidation of Fe^{++} to Fe^{+++} in sediments by overriding lava flows or by fire (e.g., an Indian campfire). I have seen cases where sediments reddened by lava flows have been confused with weathering profiles--and a case where the upper foot of sediment was blackened by being pepperitized by an overriding basalt, and the resulting profile classed as a buried Podzol by a soil scientist!

Texture is diagnostic where an appreciable concentration of clay can be discerned, called a textural B horizon, as a result of clay illuviation. The degree of clay concentration tends to increase with increase in general degree of development of the weathering profile, although it also is influenced by the amount of silt and clay in the parent material, and also by the soil-forming climate (aridity inhibits clay production and illuviation).

Clay skins (and clay coatings and clay bridges, etc.) are not only one of the most reliable criteria of a weathering profile, but also excellent indicators of the degree of illuviation.

Soil structure is influenced primarily by the amount of clay in the B horizon, particularly colloidal clay. Clay content obviously is influenced by the amount of clay present in the parent material, but for a given textural composition of parent material the clay content is proportional to the degree of soil development. Structure is a useful characteristic for identification and correlation of soils, but comparison should be made only on the basis of soils of like texture and parent material. For example, a weathering profile formed in coarse-textured material may be structureless and lack a textural B horizon, whereas in clayey material it may have well-defined blocky or prismatic structure and a textural B horizon. This is not only on account of the original clay in the clayey parent material but also because the finer grain-size of clayey soils promotes weathering and hence provides an additional amount of colloidal clay for illuviation into the B horizon.

Consistence is less reliable for differentiating B horizons of different age than color or structure. In general it is approximately the same for soils of a given age but increases with the amount of clay present. In the Kassler Quadrangle, Colo., Scott (1963) found that clayey B horizons are sticky and plastic regardless of their age, but their hardness correlates well with age. The peds in an old clayey B horizon are extremely hard; in a young clayey B horizon they are soft. As the texture becomes coarser, the B horizon becomes less sticky, less plastic, and softer, but the older soils of a particular original texture are more sticky, plastic, and hard simply because the older soils are more strongly developed and clay has had more time to form.

Thickness of the B horizon generally increases with the degree of development of the weathering profile, but it commonly differs considerably between different facies of the same geosol (see later in this chapter). The B horizon of a given facies of a geosol also tends to be thicker in coarse-grained than in fine-grained parent material. This is mainly because the higher permeability of the coarse-grained material causes illuviation to be distributed through a greater thickness, resulting in less illuviation at any one place.

Soil reaction of B horizons tends to become more acid with increasing general soil development and with increasing precipitation. Unweathered parent material is neutral or nearly so; B horizons of strongly developed weathering profiles of pedocals generally are only weakly acid, but once the pedocal-pedalfer boundary is passed,they range on pedalfers to strongly acid. The reaction of undisturbed B horizons developed in similar parent material is the same in soils of the same age and soil facies, but is more acid in the older soils.

Cca horizon.--In an area of pedocal soils the Cca (or calcic) horizon commonly is at least as distinctive as the B horizon. It also tends to be less susceptible to erosion and commonly is preserved even where the B horizon has been completely removed. A strong Cca horizon is more or less cemented (caliche) and light tan, light gray to white, commonly much lighter colored than other sediments and consequently a distinctive marker. Cca horizons of moderately and weakly developed soils are less distinctive in color contrast and cementation; nevertheless in weathered exposures they tend to develop a thin surficial crust of calcium carbonate secondarily concentrated by efflorescence. This crust is somewhat lighter colored and more resistant to erosion than the unweathered material in this horizon. It commonly causes a comparatively weak weathering profile to act as a minor caprock over less consolidated deposits. The "altithermal" soil[1/] has been an important agent in stabilizing unconsolidated late Pleistocene deposits such as dune sands throughout the semiarid western U. S., although this soil has only a moderate to weak Cca horizon.

The calcium carbonate in calcic horizons may occur in various ways (Gile, 1961): as coatings on grains, as interstice fillings, or as flakes, which may be either (1) distributed throughout the horizon, producing structures ranging from massive to blocky, laminar, platy, or bedded; or (2) segregated within the horizon, resulting in forms ranging from nodular to cylindrical, concretionary, filamentary, veined, to flaky. Cca horizons range from soft to extremely hard and from non-indurated to very strongly indurated. A distinctive feature of partly cemented gravelly calcic horizons is that the calcium carbonate coatings occur only on the lower parts of the pebbles and cobbles.

In a calcic horizon that has undergone only a single, original interval of pedogenesis the calcium carbonate is chalky white and finely crystalline,

1/ The "altithermal" soil is a widely recognized geosol in the western U. S. It formed during the later part of the "Altithermal Age" of Antevs (1948, 1952, 1955), and is represented by the Toyeh Soil in the Lake Lahontan area and the Midvale Soil in the Lake Bonneville area (see chapters 5 and 6).

commonly almost amorphous. In calcic horizons that have had secondary strong pedogenesis, however, the calcium carbonate commonly has undergone some leaching and redeposition in the upper part of the calcic horizon. This commonly results in laminated zones of nearly pure calcium carbonate, from a mere film to several inches thick, at the top of the calcic horizon, that commonly are somewhat more coarsely crystalline and locally somewhat darker than original pedogenic $CaCO_3$ concentration. Several sets of laminae may be present, resulting from multicyclic secondary deposition of calcium carbonate. Individual sets, except the youngest, commonly show truncation by younger superjacent sets; the youngest set truncates all older sets and subjacent material. In places considerable local solution and(or) fracturing of the C_{ca} horizon may intervene between the successive depositions.

Zones of secondary calcium carbonate deposition by groundwater may resemble buried calcic horizons. They commonly, though not always, occur at or near the base of a permeable zone overlying a relatively impermeable unit (e. g., in gravel overlying bedrock). Generally groundwater-deposited calcium carbonate is darker colored and more coarsely crystalline than pedogenic $CaCO_3$ (particularly that from primary pedogenesis), and it commonly (though not invariably) coats the pebbles and cobbles more or less completely, instead of just their lower parts.

Zones of cementation by lacustrine algal tufa, particularly of the "lithoid" (dense, stony) variety, also may resemble cemented calcic horizons. Lacustrine tufa typically is off-white (pale gray to pale gray-tan), whereas the calcium carbonate in C_{ca} horizons generally is whiter; lithoid lacustrine tufa commonly is somewhat more coarsely crystalline, and also commonly has concentric laminated or banded structures resulting from successive layers of algal deposition; in thin section remnants of algal filaments (thalli) commonly can be seen.

How stratigraphic position and relative age of a geosol is determined

The usefulness of a geosol as a stratigraphic marker depends on how closely its stratigraphic position (relative age) can be bracketed in terms of the Quaternary succession. Its stratigraphic-age relations can be determined by applying essentially the same stratigraphic methods and principles that are used for older rocks. Considerably more detailed work is needed, however, because of the thinness of geosols and of most surficial deposits associated with them, their lateral and vertical facies gradations, and commonly, omissions of portions of the stratigraphic sequence because of either nondeposition or erosion.

Any geosol, either relict or buried, obviously is younger than any deposits on which it has formed. Its maximum age (start of its soil-forming interval) cannot be greater than the age of the youngest underlying deposit (or unconformity).

The minimum age of a geosol (end of its interval of formation) cannot be determined accurately from relict occurrences, although in some cases an approximate minimum can be found if the age of the stability surface on which it formed is known in relation to a sequence of such surfaces (see below). Accurate determination of the minimum age can be made only from buried occurrences; the geosol is older than the oldest deposit that buries it.

Close bracketing of stratigraphic position and age, such as is necessary to meet the requirements of defining a geosol, obviously requires study of the relations of a weathering profile in a wide assortment of localities that favor fullest possible preservation and exposure of the complete depositional and soil sequence. It cannot be done in a single exposure or generally even in several exposures. Generally such determination requires detailed stratigraphic study of a comparatively large area (at least one 7½-minute quadrangle), as well as favorable expression and exposure of the stratigraphic record. Some areas (even of 15-minute quadrangle and larger size) do not reveal a stratigraphic record that has sufficient detail for defining geosols. This, however, does not preclude using stratigraphically any para-geosols that occur in these areas, if these weathering profiles can be traced into or correlated with other areas where their stratigraphic position and age has been accurately determined.

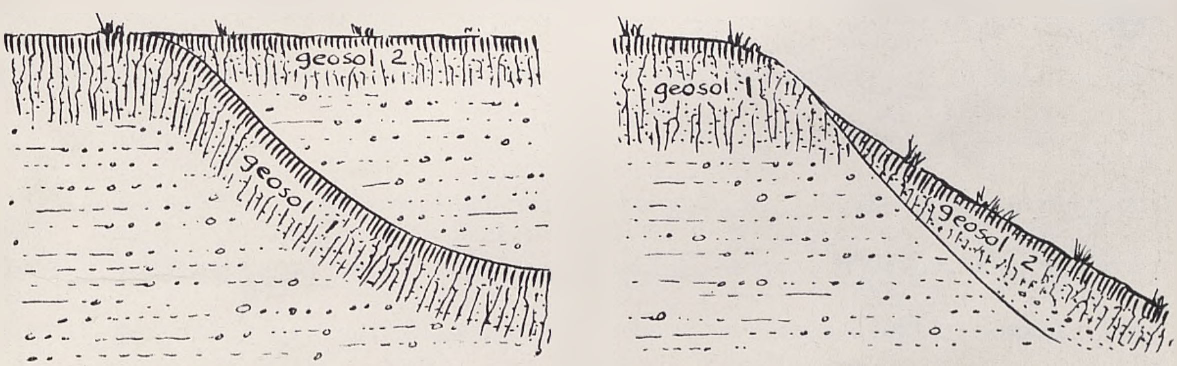
Criteria for determining the identity of a geosol

The following criteria are offered as a means of proving the identity of a given geosol--for distinguishing it from other geosols and surficial deposits. It is assumed that the geosol possesses the genetic horizons and other characteristics of a weathering profile, which, though varying, have lateral continuity.

The separate identity of the geosol can be proved by tracing it across a diversity of underlying substrates, such as surficial or bedrock deposits or other weathering profiles. Because of the discontinuity and diversity of the substrates *vis-a-vis* the relative uniformity and continuity of the goesol, none of the substrates can be part of it. If a geosol is partly eroded, its full expression can be ascertained only by tracing it laterally to a site where its full profile is preserved.

In a vertical sequence of composite or compound(ed) geosols, the identity and the vertical limits of the individual geosols can be established by means of what Butler (1959) calls the "principle of unity of profile patterns"--that is, by comparing the vertical soil-horizon patterns, which generally are repeated in their main aspects within each geosol. In the case of composite geosols, a change from one geosol to another is indicated by at least a partial repetition of horizon patterns; with compound(ed) geosols the various weathering profiles are indicated by repetition of horizon patterns at successively lower levels. In both composite and compounded sequences, contacts that are generally conformable are characteristic of the accreting zone (fig. 3.5); unconformable contacts, with one soil overlying the eroded remnants of another geosol, are characteristic of the alternating zone.

When determining the stratigraphic relations of a geosol by tracing it laterally, it is especially important to observe the character of its intersection with other geosols, and its contact relations with other deposits and erosional surfaces. One means of determining age relations of different geosols is by noting whether their intersections are what Butler (1959) calls "overplaced" or "underplaced".



A. Overplaced contact

B. Underplaced contact.

Fig. 4.1. Types of contacts between geosols

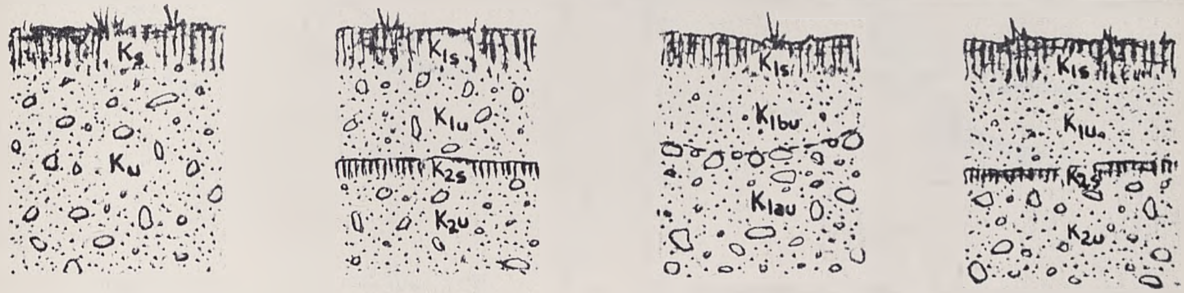
An overplaced contact (fig. 4.1,A) is where two geosols are locally separated by intervening deposits. The older geosol appears to "rise from the ground" and the younger geosol overlies it. Note that the horizons of the younger geosol are replaced from the bottom upward by those of the older geosol. Overplaced contacts obviously can be identified only where the younger geosol is weaker than the older; otherwise the relations are masked by the stronger development of the young geosol. Such contacts mark the boundary of a K-cycle unit on the depositional side, i. e., the boundary between the persistent and accreting zones.

An underplaced contact is that where development of an erosion surface has intervened between formation of two geosols (fig. 4.1,B). The older geosol appears to "run off into the air", i. e., is cut off by the erosion surface, and the younger geosol comes in below it. Note that the horizons of the older geosol are replaced from the top down by those of the younger geosol. Underplaced contacts generally mark the boundary between the sloughing zone and the persistent zone of K-cycle units. The angle of the underplaced contact commonly is influenced by the hardness of the older geosol relative to the underlying material. Where the older geosol has hard horizons over softer ones or soft parent material, a scarp tends to develop at the contact.

In depositional (accreting) zones the younger geosols successively overlie the earlier ones; in erosional (sloughing) zones, however, the younger geosols generally are cut into the older ones and consequently usually lie topographically below the older ones. Unpredictable relationships in levels occur where deposition has followed cutting, i.e., in the alternating zone, but in all cases the younger geosols are inlaid into the older.

Typical examples of how stratigraphic and age relations of geosols are determined

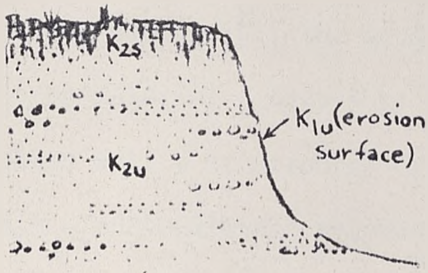
The following examples show the usual kinds of relationships from which the stratigraphic position and relative age of Quaternary geosols can be determined.



A	B	C	D
Deposition, then stability (1 K cycle)	Deposition, then stability; 2d deposition, then 2nd stability (2 K cycles)	Deposition, then different-type deposition, then stability (1 K cycle)	Deposition, then stability; 2d (different) deposition, then 2nd stability (2 K cycles)

Fig. 4.2. Basic instability-stability (weathering) relations

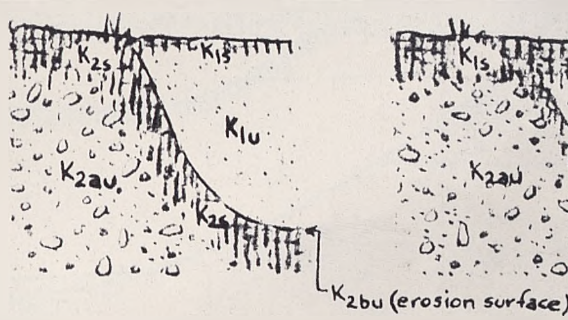
Simple vertical relations are shown in the above figure. Part A records a single erosion-weathering cycle (K cycle). Deposit K_u formed during its instability interval, and weathering profile K_s during its stability interval. Part B shows two K cycles, with an early instability deposit (K_{2u}) overlain by an early geosol (K_{2s}, now buried), and then by a similar younger deposit (K_{1u}) and then another geosol (K_{1s}). Part C again records a single K cycle, but with change in type of instability deposit with time. K_{1au} is the earlier facies and K_{1bu} the later facies. In this type of relationship it is important to check carefully, by lateral tracing if necessary, whether a soil actually may lie stratigraphically between K_{1au} and K_{1bu}, being only locally eroded. Part D shows such a case, which records two K cycles, separated by such a buried geosol and with the younger deposit of different type than the earlier.



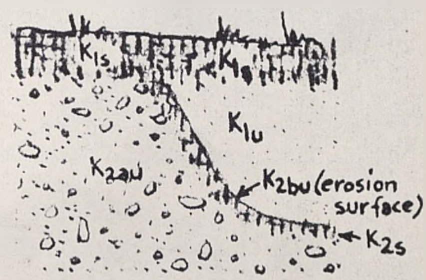
A
Deposition, then stability, then erosion (2 K cycles)

B
Deposition, then erosion, then stability (1 K cycle)

C
Deposition, then erosion; 2d deposition, then stability (1 K cycle)



D
Deposition, then erosion, then stability (strong weathering) then 2d deposition, 2d stability (weak weathering)(2 K cycles)



E
Deposition, then erosion, stability (weak weathering); 2d deposition, 2d stability (strong weathering)(2 K cycles)

Fig. 4.3. Typical relations between geosols and erosion surfaces

Typical geosol-erosion surface-deposition relations are shown above. In A, the deposit and geosol of an early K cycle have been dissected during the instability phase of a later K cycle, which has formed the K_{1u} erosion surface. In B, however, only one K cycle is recorded, with erosion intervening between the depositional and stability phases. In C, two successive phases of differing deposition (with erosion intervening between or synchronous with the second deposition) are recorded prior to stability, all in one K cycle. D records two K cycles, the first one of the same kind as shown in B, ending in strong weathering, the second K cycle consisting of deposition of different-type material followed by weak weathering, which has not megascopically affected relict occurrences of the older geosol. E records the same kind of erosional-depositional sequence, except the older weathering profile is weakly, and the younger one, strongly developed. The younger geosol has completely masked relict occurrences of the older weathering profile.

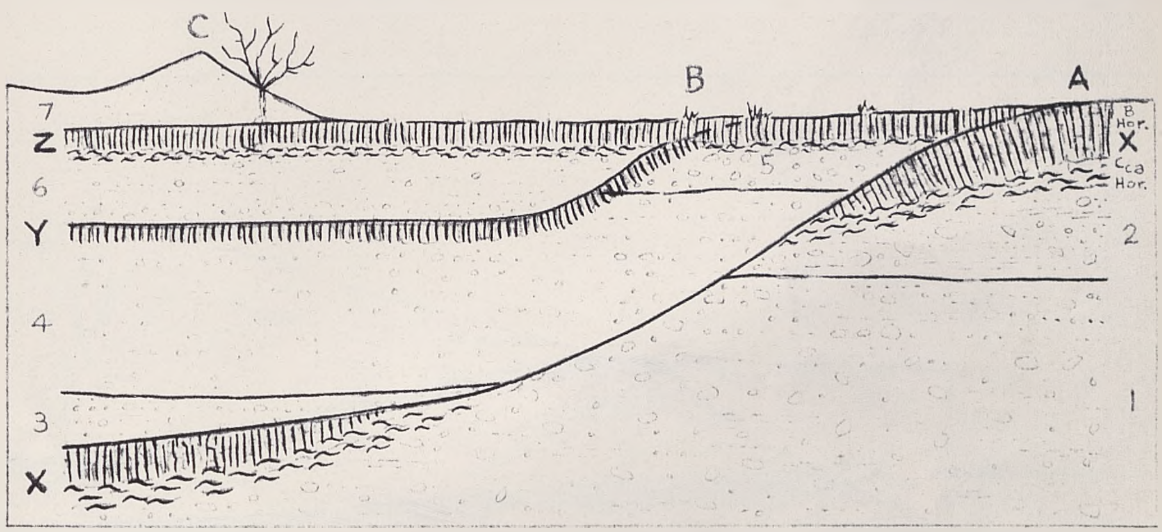


Fig. 4.4. How stratigraphic position of geosols can be determined by lateral tracing.

Fig. 4.4 illustrates a typical instance of how the stratigraphic position of geosols can be identified by lateral tracing. At A, only the very strongly developed geosol X can be identified, in a relict occurrence, and can be seen to be developed on deposits as young as unit 2. By tracing leftward, this geosol can be seen to be overlain unconformably (with an overplaced contact) by the relict, moderately developed geosol Z. At B, only geosol Z can be identified, developed on deposits as young as unit 5; geosol X is here locally eroded by the erosion surface that is now known to be younger than geosol X and older than unit 4. By tracing still further leftward, still another geosol Y, can be identified, developed on deposits as young as unit 5. This geosol is preserved only where buried because it is more weakly developed than geosol Z and in relict occurrences it is completely masked by the latter. In the vicinity of C, geosol Y can be seen to be overlain by deposits as old as unit 6. Likewise, geosol Z overlies unit 6 and is overlain by unit 7. In conclusion, the stratigraphic position of geosol X can be bracketed between units 2 and 3; of geosol Y, between units 5 and 6; and of geosol Z, between units 6 and 7.

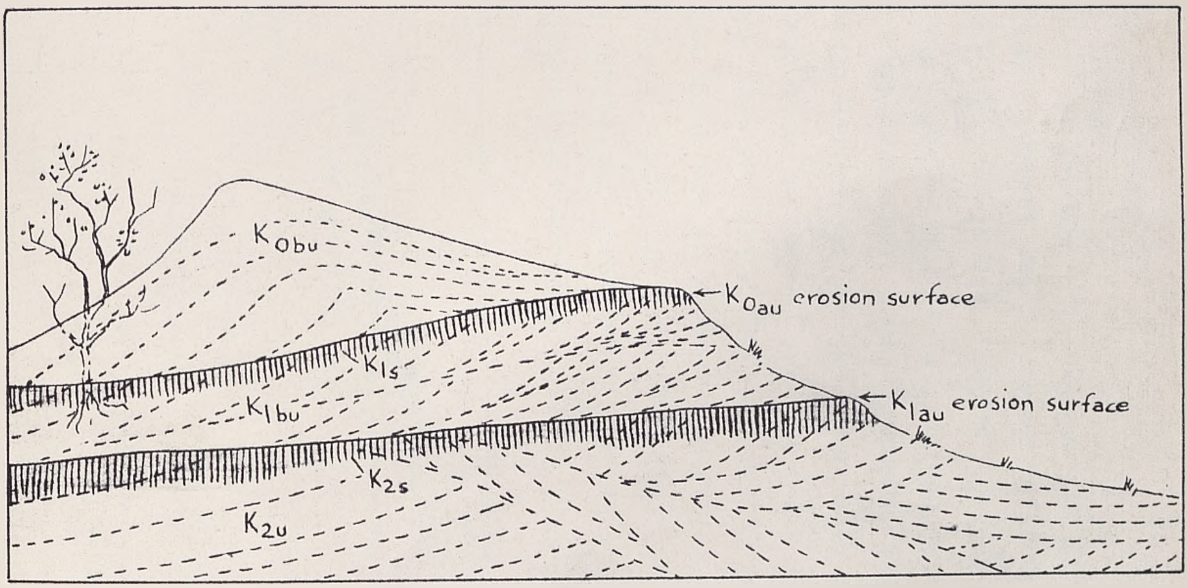


Fig. 4.5. Typical relations in areas of eolian sand deposition and erosion.

Fig. 4.5 shows relations typical of landscapes of eolian deposition and erosion. An early erosion-weathering (K) cycle is recorded by eolian sand unit K_{2u} and by geosol K_{2s}. The next K cycle is recorded by sequence: K_{1au} erosion surface, K_{1bu} eolian sand, and finally, the K_{1s} geosol. The present (incomplete) K cycle is recorded by the K_{0au} erosion surface and the overlying active eolian sand unit K_{0bu}.

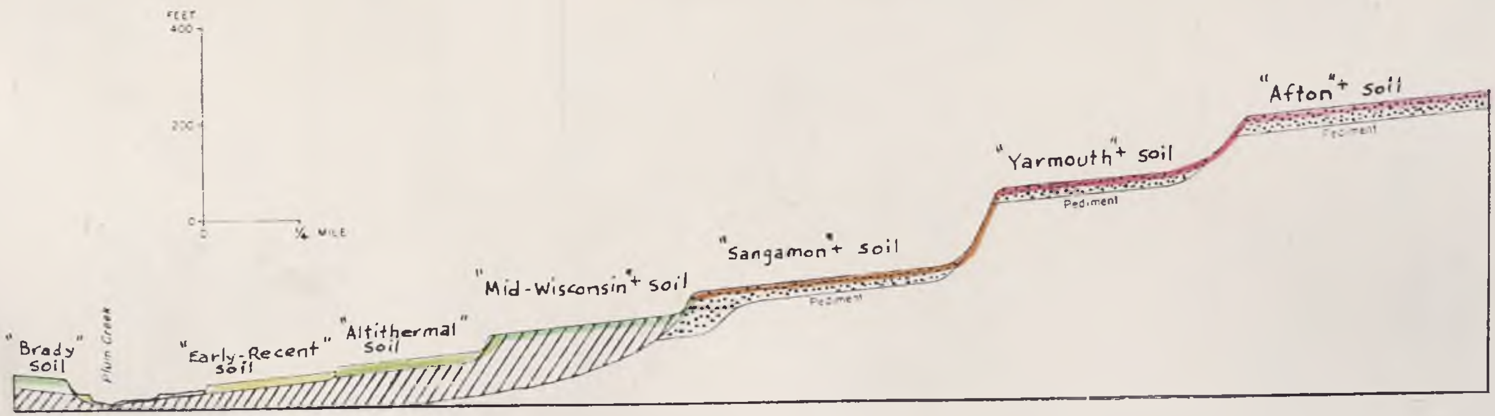


Fig. 4.6. Typical topographic relations of relict occurrences of Quaternary geosols (and weathering profiles) on pediment and stream-terrace gravels. In such a sequence, the relict geosols on higher pediments and terraces normally are older than those on lower terraces. The cross section shows diagrammatically the relations at the western boundary of the Colorado Piedmont section of the Great Plains, near Denver, Colorado (modified after Scott, 1963). The Afton, Yarmouth, and Sangamon soils are very strongly developed relict weathering profiles on the stability surfaces of the three highest and oldest Quaternary alluvial units, which veneer pediments. The plus (+) symbol after the soil names denotes that these relict weathering profiles record the summation of pedogenic and other surficial processes since their respective original pedogenic intervals. The Sangamon+ profile usually is distinctive and qualifies as a geosol. The Yarmouth+ profile is even more strongly developed and also qualifies as a geosol (on the basis of specific identity) in most places. The Afton+ weathering profile, however, typically is so masked by the Yarmouth weathering profile that it cannot be separately identified, and it qualifies only as a weathering profile.

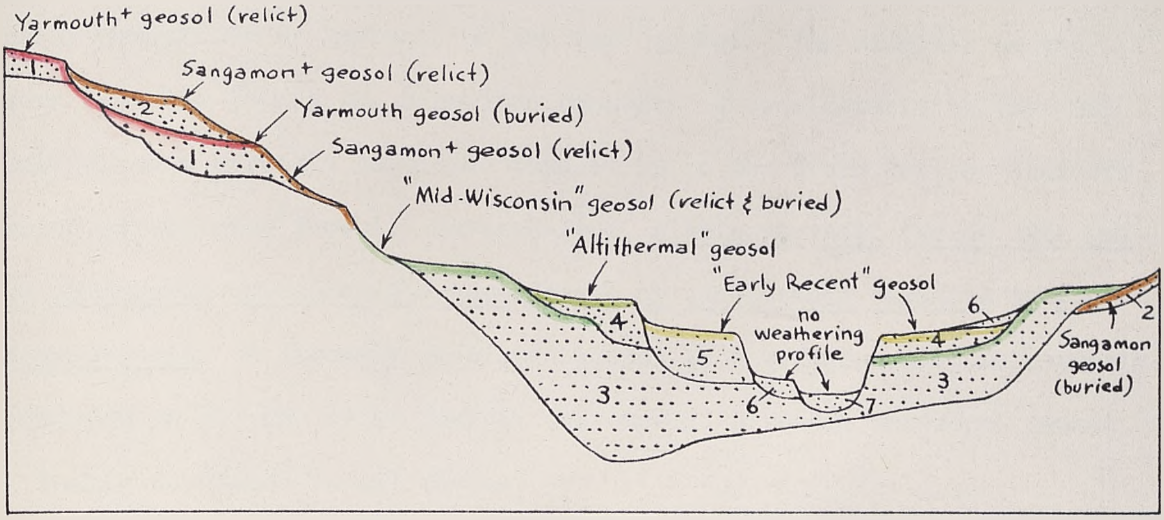


Fig. 4.7. Typical geosol and alluvial cut-and-fill relations along western margin of the Great Plains.

The above diagrammatic cross section shows geosol and alluvial cut-and-fill relations typical of the Colorado Piedmont section of the Great Plains near Denver, Colorado. The successive alluvial units (locally with colluvium) are numbered from 1 (oldest) to 6 (youngest).

Problems in use of geosols as stratigraphic markers

Problems of two chief kinds arise in identifying geosols for use as stratigraphic markers consistently throughout an area: (1) Soil facies differences--differences in the original type and degree of development of a given geosol because of variations in local environmental conditions that existed at the time the geosol was forming; (2) secondary modification of the geosol after it has formed, by geologic, climatic, and biologic factors.

Identification and correlation problems caused by soil facies

Separate facies of a geosol may differ considerably in key characteristics such as thickness, color, structure, consistence, and reaction of the B and Cca horizons. Where the facies differences are large, as on opposite sides of the pedalfer-pedocal boundary, the separate facies may not be easily recognized as variants of the same geosol. For example, in the Brown Podzolic facies the B horizon commonly is considerably thicker and somewhat more strongly developed than it is in the Brown Soil or Sierozem facies of the same geosol. Nevertheless, <u>within any soil facies each geosol invariably maintains the same degree of development relative to the other geosols</u>. A strongly developed geosol in one facies also is strongly developed in all the other facies; conversely, a weakly developed geosol is weakly developed in all facies.

Where local environmental factors cause significant variations in geosols, an effort should be made to select areas where only one or two of the soil-forming factors differ, thus giving some basis for estimating their relative influence. Thus, in determining the relative degree of development of various geosols in a stratigraphic sequence, it is highly desirable to compare the soil characteristics in well-drained profiles on flat or gentle slopes at similar altitudes and on unconsolidated parent materials of similar texture, and mineralogic composition. Particularly to be avoided is comparison of profiles from markedly different altitudes, especially if they are on opposite sides of a pedocal-pedalfer facies boundary, from both well-drained and poorly drained locations, and from parent material that differs considerably in texture or in mineralogic composition--such soils are very difficult or impossible to compare on a common basis. Composite or superposed soil profiles, recording the effects of more than one soil-forming optimum, should be avoided wherever possible.

In summary, in order reliably to use geosols for stratigraphic differentiation and correlation throughout an area with considerable differences in climate, vegetation, parent material, drainage, and relief, it is necessary to determine the typical development of each geosol in each principal soil facies, and then to compare the geosols on the basis of their relative degrees of development and stratigraphic association within each facies. If this is not done there is danger of miscorrelating geosols between different facies. It is particularly helpful to trace the geosols across soil-facies boundaries, either by means of exposures of buried occurrences or by means of relict occurrences on landforms or geomorphic surfaces of known age. Such opportunities should be sought for, because they afford the surest means of perfecting the correlations of geosols throughout any area that has markedly differing soil facies.

Some general deductions on the climatic-vegetation factors as facies determinants.--Climate and vegetation (the principal part of the organisms soil-forming factor) will be considered together because climate is the main determinant of plant ecology. Certain general deductions can be drawn from Quaternary soil-stratigraphic studies in the western U. S.

1) In any sizeable area, particularly one with varied terrain, the altitude of the boundary between any two adjacent climatically-determined soil facies will vary by several hundred feet, depending on local (mostly microclimatic) environmental variables, such as north vs. south-facing slopes.

2) The stronger the development of a geosol the higher the average altitude of all its climatically-determined soil-facies boundaries. For example, in the La Sal Mountains, Utah, (Richmond, 1962a, p. 38, fig. 19) the average altitude of the pedocal-pedalfer facies boundary (between the Brown Forest and Brown Podzolic soil facies) is about 8900 ft. for the very strongly developed geosols of pre-Wisconsinan age, 8700 ft. for the strongly developed geosol of middle Wisconsinan age, 8000 ft. for the moderately developed geosol of "altithermal" age, and only 7000 ft. for two weakly developed soils (fig. 48). Likewise, comparable variation with altitude was found for

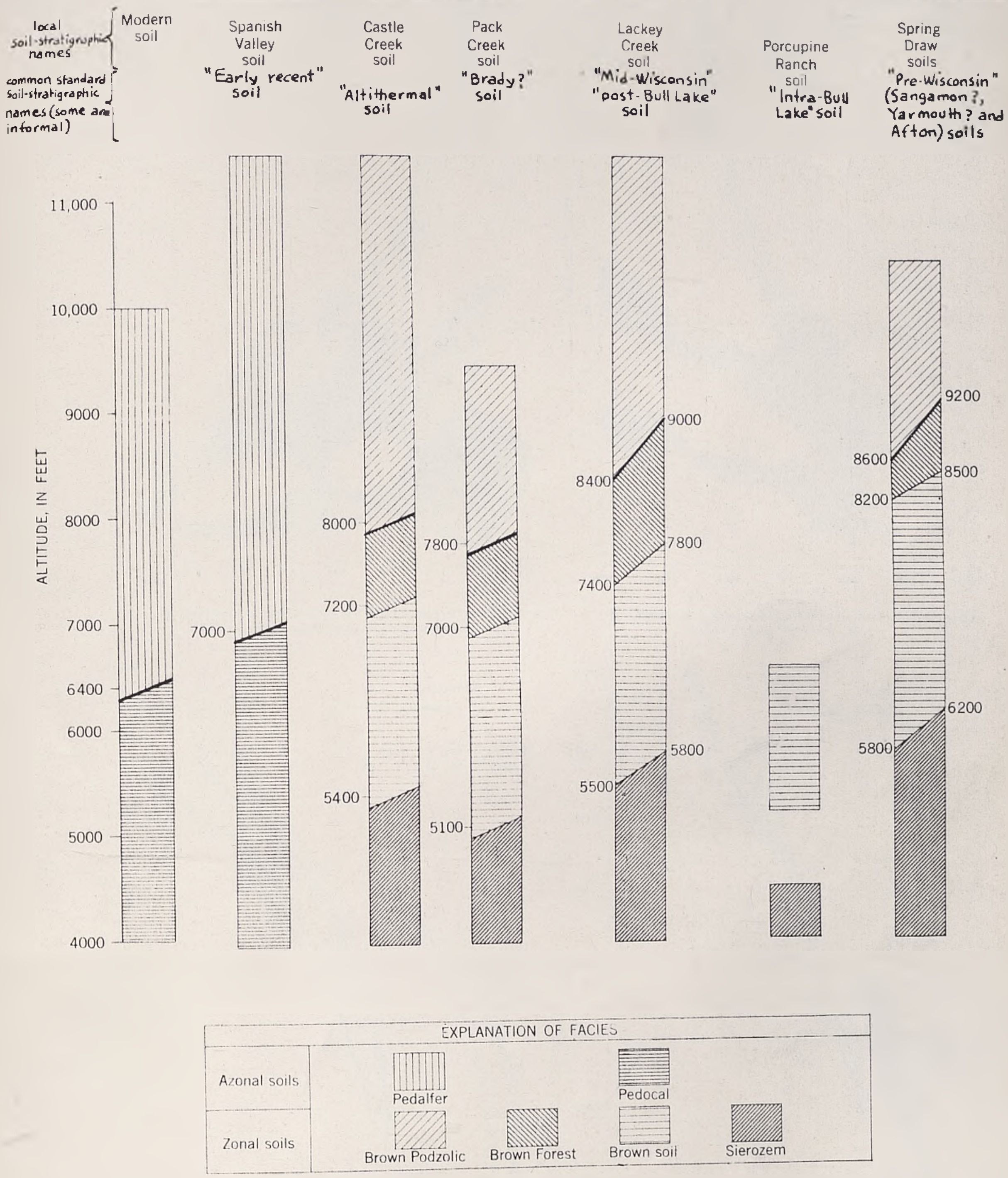
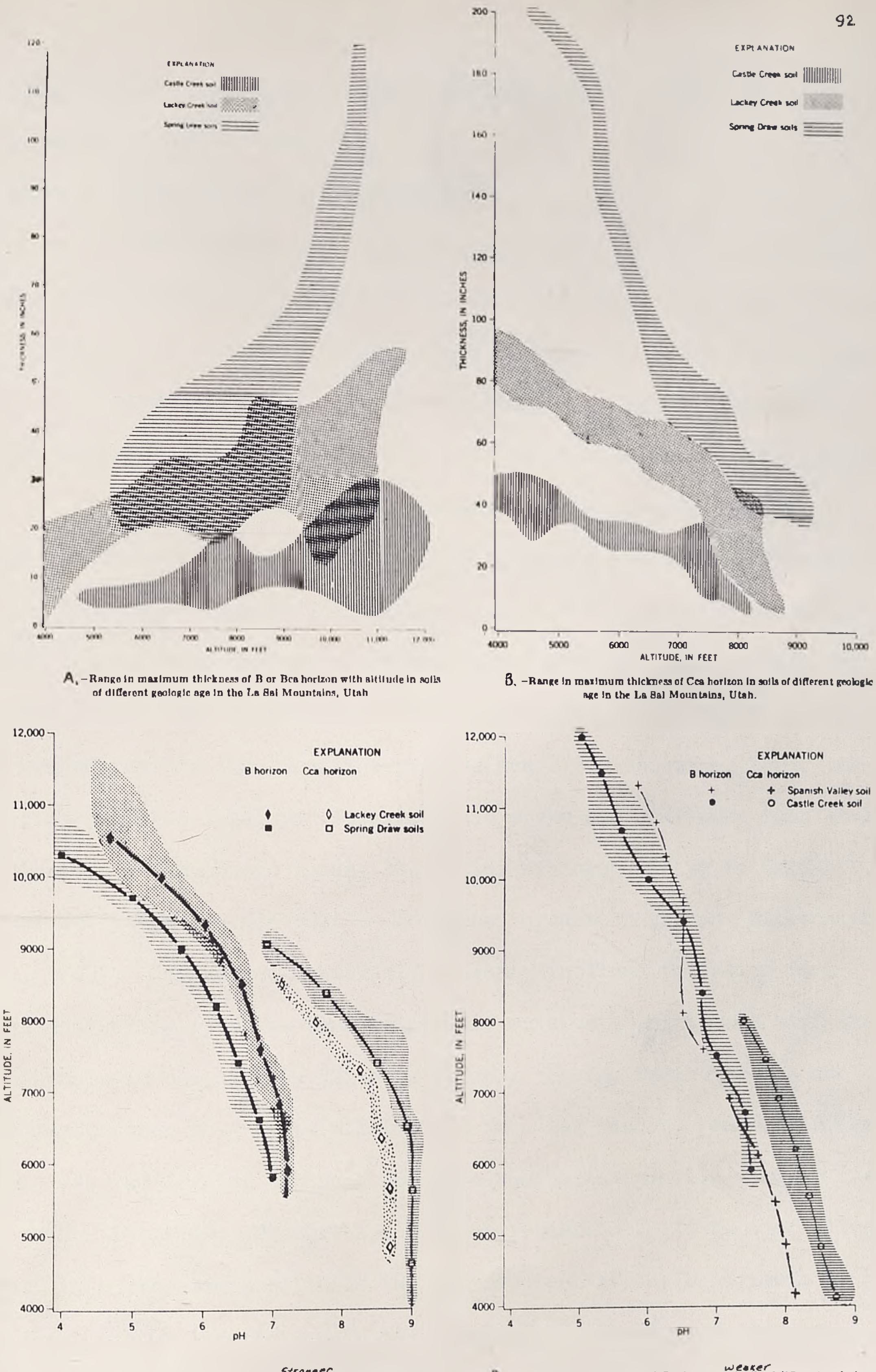


Fig. 4.8. Range in altitude of facies and facies changes of geosols of different geologic age in the La Sal Mountains, Utah. Altitudes beside columns are average altitudes of facies changes. (After Richmond, 1962a)



A.–Range in maximum thickness of B or Bca horizon with altitude in soils of different geologic age in the La Sal Mountains, Utah

B.–Range in maximum thickness of Cca horizon in soils of different geologic age in the La Sal Mountains, Utah.

C.–Distribution of pH of B and Cca horizons of soils of different geologic age with altitude in the La Sal Mountains, Utah.

D.–Distribution of pH of B and Cca horizons of soils of different geologic age with altitude in the La Sal Mountains, Utah.

Fig. 4.9. Graphs showing changes in properties of several geosols with altitude in the La Sal Mountains, Utah. A, range in maximum thickness of B or Bca horizon; B, range in maximum thickness of Cca horizon; C and D, range in pH of B and Cca horizons. (The Spring Draw soils are of pre-Wisconsinan age and are very strongly developed; the Lackey Creek soil is of mid-Wisconsinan age and strongly developed; the Castle Creek soil is of altithermal age and moderately developed; and the Spanish Valley soil is of early Recent age and weakly developed.) (Modified after Richmond, 1962a.)

properties such as soil reaction (pH), thickness of B or Bca horizons, and thickness of Cca horizons (fig. 4.9).

General relations of the parent material factor.--A complete description of soil parent material requires statement of the mineralogical and chemical composition, the texture (size distribution) of the component particles (whether of an unconsolidated granular material or of a crystalline igneous or metamorphic rock), and the structural arrangement (including bedding, jointing, schistosity, dip of bedding or schistosity, permeability, etc.). Each of these variables potentially can exert a powerful influence on the type of weathering profile that is developed. For example, in humid temperate climates the weathering profiles on erosion surfaces of the same age are much deeper on coarse-grained than on fine-grained igneous rocks of the same composition; obviously the coarse-textured rocks weather more rapidly than the fine-grained ones. On the other hand, the weathering profiles developed from granitic rocks are shallower than those from dioritic and gabbroic rocks; the latter are also richer in Ca and P and have mull-type (in contrast to mor-type) organic horizons (Muckenhirn et al., 1949). Gneiss and schist with steeply inclined foliation form thicker weathering profiles than relatively massive granitic rocks of the same mineralogic composition (Hilgard, 1906). In the humid tropics obsidian may weather only a few millimeters in a few hundred years, whereas volcanic ash may develop a strong podzolic weathering profile several feet thick (Mohr, 1944; Geo. Frazer, U.S.G.S., oral commun.). In further illustration, in northeastern Illinois the clay content of the B horizons of soils developed on late Wisconsin till correlates closely with the original clay content of the till (Stauffer, 1935), as also does the depth of leaching of calcium carbonate, which decreases as the clay content increases.

In general, however, the effect of the parent material on a weathering profile tends to decrease as the soil-development increases. This means that strongly and very strongly developed weathering profiles tend to be similar, in spite of large differences in parent material; conversely, weak weathering profiles may differ considerably with comparatively small variations in parent material. The influence of parent material tends to persist longer in semiarid and arid regions than in humid ones, because hydrolysis, leaching and other chemical weathering processes progress more slowly.

The effect of parent material is at its maximum for extremes of textural and chemical composition. The texture of the parent material, especially its permeability, strongly influences the depth (thickness) of the whole soil profile, and particularly that of the B (and Cca, Csi, etc., if present) horizons. The thickness is greater on light- than on heavy-textured parent materials.

The influence of compositional variations in parent material is discussed comprehensively by Polynov (1930), and Stauffer (1935) has discussed the relations in a humid temperate climate. Chemical compositional effects are pronounced where certain elements that normally occur in parent materials are markedly deficient, or where certain elements occur in unusual proportions and remain insoluble and become concentrated on weathering. The most common example is limestone; indeed all calcareous parent materials have unique properties under pedogenesis. Carbonates of Ca and Mg are readily dissolved and carried away in humid regions, and they are leached from the upper part of the weathering profile in all but very arid climates. In humid regions, however, alkaline earth carbonates retard the podzolic-type weathering processes. Colloidal sols of $Al(OH)_3$ and $Fe(OH)_3$, where unprotected by colloidal humus, became readily flocculated by calcium carbonate, thus reducing translocation of colloidal clay particles and impeding the eluviation-illuviation process. Residual soils derived from phosphatic limestone commonly remain high in phosphorous because this element is not readily leached and tends to be concentrated in the weathering profile, even where the calcium carbonate is completely leached.

Striking examples of deficiency of alkaline earths locally can be seen in pedocal terrains underlain by volcanic rocks of intermediate composition (e.g., andesite or dacite), as in many parts of the Basin-and-Range Province: usually the weathering profiles have strong Cca horizons, but in sites where hydrothermal alteration has locally removed the calcium and magnesium the soils have only weak or no Cca horizons.

The relief-drainage factor.--Relief (slope) influences surface and subsurface drainage, and consequently the amount of water that enters the soil profile, and in turn the extent and type of profile development. Relief therefore helps to determine such soil properties as: (a) coloration and degree of mottling (gleying), (b) thickness and organic matter content of the surface horizon, (c) depth of solum, (d) degree of horizon differentiation, (e) salinity, and (f) kind of hardpan. Relief can be said to act akin to local microclimatic differences: convex uplands readily lose runoff and tend to be drier, whereas concave slopes tend to receive runoff (fig. 4.1.1). As the slope decreases or changes from convex to concave, the soils tend to become increasingly more fine-textured, darker, with more organic matter andd thicker surface horizons; they may also show gleying or other effects of an intermittently or permanently high watertable, and in arid regions they may become saline (figs. 4.1.2; 4.1.3; 4.1.4).

In general, the effects of relief and drainage are lessened with aridity, so that in many semiarid and arid regions they are negligible; conversely, they may be relatively important in humid regions. In part, however, this is because of the greater influence of mass-waste processes such as solifuction in humid areas.

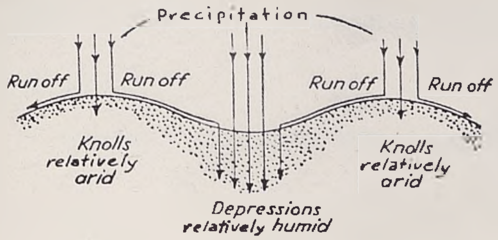


Fig. 4.1.1. Effect of relief on runoff and water penetration of soils (after Jenny, 1941).

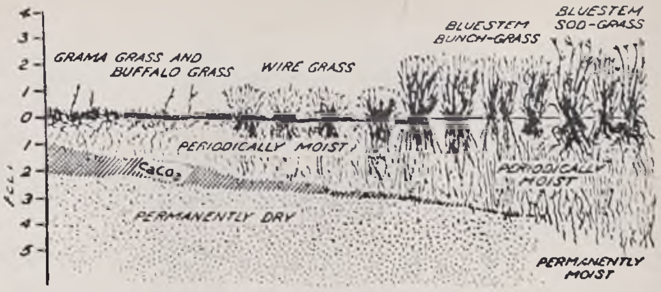


Fig. 4.1.2. Diagram showing relation of plant communities to the depth of penetration and amount of soil moisture in the Great Plains area. (After Jenny,1941.)

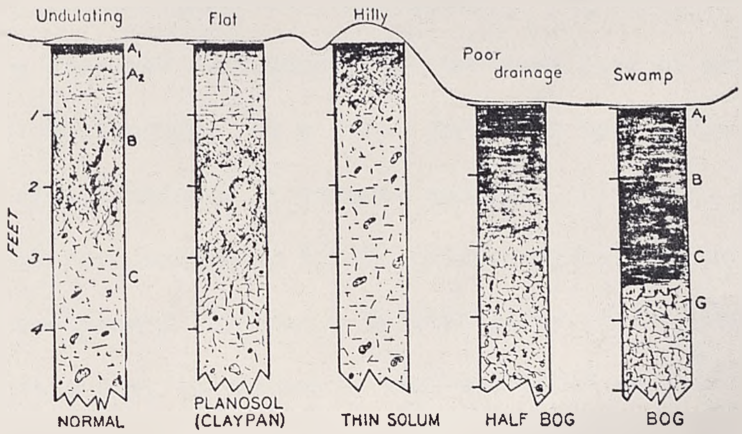


Fig. 4.1.3. Typical changes in soil facies because of variations in relief and drainage. (The profiles are all of the same age and on similar parent material.) Note that a shallow profile is developed on hilly land because of high runoff and erosion, wheras a thick profile is developed on the flat upland because of little or no erosion, and this profile is highly leached in its upper part and has a dense claypan in its lower part. (After U. S. Dept. of Agriculture, 1938)

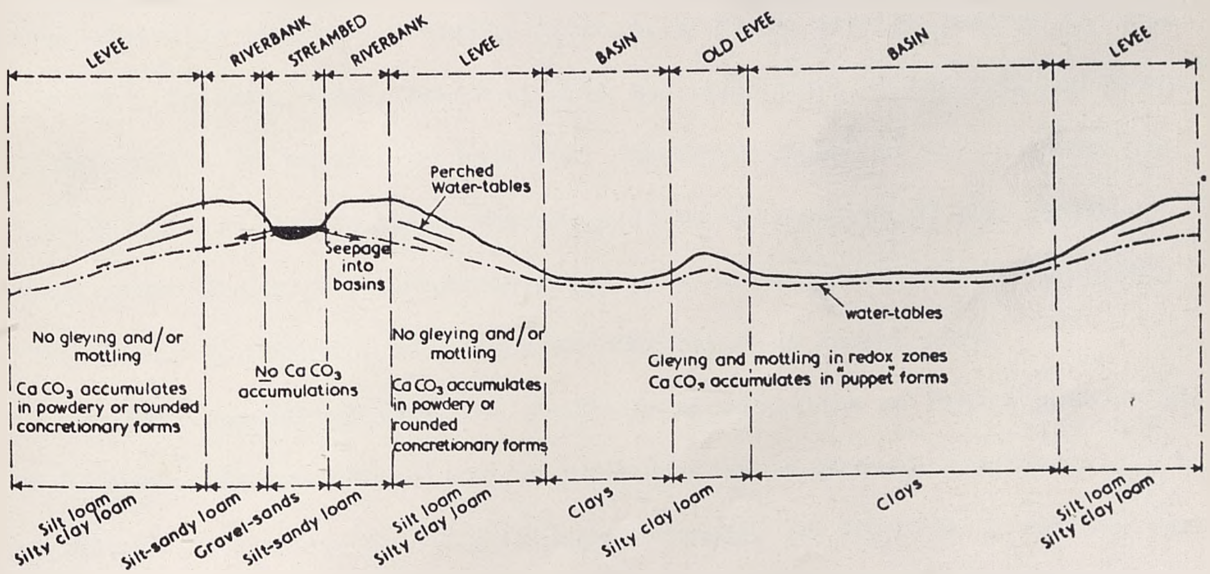


Fig. 4.1.4. Diagrammatic cross section showing changes in soil facies because of variations in surface relief and drainage in an arid climate, and particularly characteristic of irrigated areas. Typical of conditions on the Mazanderan Plain, Iran. (After Andriesse, 1960).

Identification-correlation problems caused by secondary modifications of geosols

Geosols may have their profiles modified by various subsequent changes, in either buried or relict occurrences.

Relict occurrences.--Relict geosols are most liable to secondary modification, inasmuch as they remain exposed at the land surface and vulnerable to the continuing surficial environment. Very few, if any, surficial processes have operated continuously at the same rate at any given locality, throughout the Quaternary. As a rule, there has been a discontinuous succession of periodic changes in various geologic, biologic, and pedogenic processes, that were triggered by the cyclic climatic changes that characterize the Quaternary (typically in most parts of the western U. S. from cool-dry, to warm-dry, to warm-wet, cool-wet, cold-wet, and back to cool-dry). These climatic changes in turn produced changes in rate and even in process-type of biotic activity (especially in vegetation), of physico-chemical mineral weathering activity, and of geologic erosional-depositional activity (fig. 3.4, table 4.2). The subsequent activity may be so great that the original geosol is completely masked or lost, as where erosion removes it completely. In many cases, however, only the upper part of the original weathering profile is removed or modified. Wherever enough of the original profile is preserved to display characteristics distinctive of this geosol, these truncated relict occurrences can be used for stratigraphic differentiation and correlation with as much confidence as buried ones.

A few examples of kinds of secondary modification of relict geosols are as follows: Truncation of the original profile by geologic erosion (e.g., by deflation, slopewash, or solifuction); intrusion or stirring of surface (generally organic A) horizon material into the former B horizon by activity of burrowing animals such as ground squirrels, earthworms, by deeply penetrating

Table 4.2

Possible kinds of secondary modification of relict geosols as a result of climatic changes and resulting geologic and pedogenic changes. (The list is not intended to be comprehensive.)

Climatic changes: Temperature-precipitation intervals in typical glacial-interglacial (pluvial-interpluvial) cycle in the western U. S.				Effect of the climatic change on an existing geosol and on the geologic surficial environment. (Note: These effects vary with latitude, altitude, and other aspects of the general climatic situation; the effects given are for the coterminous western U. S. between 3,000 and 9,000 ft. altitude.)
TIME	Inter-glacial interval	Pluvial interval	Warm-wet	End of original soil-forming interval
TIME	Inter-glacial interval	Pluvial interval	Cool-wet	Beginning of instability interval: Active erosion by slopewash, stream degradation, mass-wasting (solifluction, creep, landsliding, etc.), and resultant local deposition.
TIME	Glacial interval	Pluvial interval	Cold-wet	Maximum instability in most areas: Active fluvial and mass-waste erosion and deposition; frost action (especially in periglacial areas); local glaciation and high stages of pluvial lakes.
TIME	Glacial interval	Interpluvial interval	Cold-dry	Waning of instability interval: Lessened stream and mass-waste erosion and deposition (alluvium deposited becomes scantier and finer); waning glaciation, frost action, and pluvial lakes.
TIME	Interglacial interval	Interpluvial interval	Cool-dry / Warm-dry	Secondary maximum (local) of instability: Eolian activity (deflation, deposition of loess and eolian sand); local slopewash erosion and deposition and arroyo cutting.
TIME	Interglacial interval	Pluvial interval	Warm-wet	Stability interval: Development of next weathering profile. Near-cessation of geologic erosion and deposition in most areas.

Other possible geologic modifications: Churning of surface and(or) deeper soil horizons (and commonly also upward movement of pebbles and stones) because of alternate deep cracking, sloughing, and closing of cracks consequent upon alternate drying and wetting, or because of frost-stirring (congeliturbation).

roots, etc.; development of a secondary calcic horizon in the original B horizon of a strongly developed pedocal by exposure to a subsequent soil-forming interval. [The latter is particularly liable to occur if a thin layer of new parent material is deposited over the former weathering profile, by wind (as loess) or by water (slopewash)]

Preservation of relict geosols is favored by vegetative cover, by coarse gravelly parent material, by flat or very gentle slopes, and by absence of torrential rains and strong frost action. In contrast, the older relict geosols commonly are completely stripped or otherwise modified where they were developed in desert areas characterized by severe torrential rains and without adequate vegetative cover; likewise, in cold wet regions they may be more or less completely reworked by frost stirring.

Buried occurrences.--A buried geosol may undergo secondary modification prior to burial, while it still is exposed as a relict soil profile. For example, many geosols buried by till or by lava show shove effects caused by the overriding glacial ice or lava. Buried geosols also may undergo chemical and physical modification by subsurface agencies, such as compaction and induration, or deposition or leaching of calcium carbonate, other salts, iron or manganese hydrous oxides, etc. by seeping and percolating groundwater. Organic A horizons are much less persistent (resistant to modification) than B horizons resulting from gain or loss of colloidal clay and/or sesquioxides; cemented horizons are said to persist indefinitely, and well-developed blocky or prismatic soil structure also is very persistent.

CHAPTER V

EXAMPLES OF MODERN STRATIGRAPHIC STUDIES OF QUATERNARY DEPOSITS USING WEATHERING PROFILES (GEOSOLS) AS SOIL-STRATIGRAPHIC UNITS

Modern stratigraphic studies of surficial deposits in various parts of the U. S. have shown that the Quaternary successions invariably contain a sequence of weathering profiles (soils), and that the age of a given weathering profile can be determined in relation to the sedimentary sequence. Commonly the stronger weathering profiles are the most reliable rock-stratigraphic markers that facilitate subdivision and mapping of units on a local rock-stratigraphic basis. The soil sequences have been found to be remarkably similar from area to area, in terms of relative age and relative development of the weathering profiles.

Three areas have been selected as examples of those in which modern soil-stratigraphic-oriented studies of surficial deposits have been made. The Quaternary successions in these areas are not only well preserved, well exposed, and well studied, but also they have what might be termed "lithologic character", in that they are sufficiently varied as to give unusually sensitive and detailed records of varying types and rates of deposition and erosion, induced by the Quaternary climatic changes. These three areas are by no means unique; the list of examples of excellent studies and rewarding areas could be expanded considerably.

Lake Lahontan area, Nevada

The Quaternary deposits of the Lake Lahontan area, Nevada, have been studied by Morrison (1952a, 1952b, 1961c, 1961d, in press, a). They comprise seven main units (table 5.1), which are, from oldest to youngest:

1. Lacustrine sediments of pre-Lake Lahontan age.

Table 5.1. Late Quaternary stratigraphy and lake history in the southern Carson Desert, near Fallon, Nevada

Period	Epoch	Time		Rock- and Soil-Stratigraphic units	Description	Max. Exposed thickness (ft.)	Exposed Altitude range (ft.)	Lake history recorded*
Quaternary period	Recent epoch	post-Lake Lahontan time	present Late post-Lake Lahontan shallow-lake and desiccation interval	Fallon Formation	Eolian sand and alluvium intertonguing with shallow-lake sand to clay and tufa.	10	above 3865	2 main lake cycles, preceded, separated, and followed by desiccations; lake maxima at 3919 ft.(both cycles) in Carson Lake area, and 3905 and 3890 ft.in northern area.
					Shallow-lake sand to clay, minor tufa.	5	3870-3922	Third post-Lahontan lake cycle, max. at 3922 ft.in Carson Lake area, 3915 ft. in northern area.
					Eolian sand and alluvial sand and silt; very weak soil (L Drain Soil)	5	above 3880	Desiccation, essentially complete.
					Shallow-lake sand to clay, minor tufa.	6	3870-3930	Second post-Lahontan lake cycle, max. at 3930 ft.
					Eolian sand, alluvial sand and silt.	4	above 3900	Desiccation, probably nearly complete.
					Shallow-lake sand to clay, minor gravel and tufa.	6	3870-3950	First post-Lahontan lake cycle, max. at 3950 ft.
	Pleistocene epoch		Early post-Lake Lahontan desiccation interval	Toyeh Soil	Gray Desert soil, moderately developed; widespread.	1.5	above 3880	Major post-Lake Lahontan desiccation, complete.
				Turupah Fm.	Eolian sand, widespread, and alluvial sand and gravel, local. Contemporaneous disconformity records deflation as low as 3865 ft.	30	above 3875	
		Lake Lahontan time	Late Lake Lahontan deep-lake interval	Sehoo Formation (lacustrine) and Indian Lakes Formation (subaerial)	Lacustrine sand to clay, minor gravel and tufa; fairly widespread.	16	3875-3990	Fifth Lahontan lake cycle, maximum at 3990 ft.
					Colluvial sand to clay; swamp muck; bears very weak soil (Harmon School Soil)	3	above 3900	Temporary lake recession, to at least 3900 ft.
					Deep-lake gravel to clay, dendritic and lithoid tufa, widespread.	25	3865-4250	Fourth Lahontan lake cycle, maximum at 4190 ft.
					Alluvial sand and gravel, eolian sand, very weak soil, all (rare); minor thinolite tufa, lake silt and clay.	5	Subaerial sediments above 3990; lake sediments, 3890-4000	Temporary lake recession, to at least 3990 ft.
					Deep-lake gravel to clay, lithoid, "cellular", and "coralline" tufa, widespread.	50	3865-4370	Third Lahontan lake cycle; maximum at 4370 ft.
			Mid-Lake Lahontan desiccation interval	Churchill Soil	Brown soil, strongly developed, locally preserved.	3	above 3920	Major lake recession and intermittent desiccation; lakes were shallow and temporary.
				Wyemaha Fm.	Eolian sand and alluvial sand and gravel, intertonguing with shallow-lake sand to clay.	100	subaerial sediments above 3880; lake sediments, 3865-3990	
			Early Lake Lahontan deep-lake interval	Eetza Formation	Deep-lake gravel, minor sand to clay and tufa; local.	40	3940-4340	Second Lahontan lake cycle, maximum at about 4340 ft.
					Colluvial and alluvial gravel and sand; very weak soil; (unnamed).	3	above 4200	Temporary lake recession; exposures along lower Truckee R. above Nixon suggest lake fell at least to 4100 ft.
					Deep-lake gravel, minor lake sand to clay and tufa, local.	50	3950-4380	First Lahontan lake cycle; maximum at 4380 ft. is highest of Lake Lahontan.
			Late pre-Lake Lahontan recession interval	Cocoon Soil	Brown soil, very strongly developed, heavy Cca (caliche) horizon; locally preserved.	4.5	above 3960	Major lake recession or desiccation; no lakes above 3960 ft.
				Paiute Fm.	Alluvial and colluvial gravel, local.	40	above 3960	
			Pre-Lake Lahontan deep-lake interval	Upper pre-Lake Lahontan lacustrine unit	Lacustrine gravel and sand.	20	4020-4100	Deep lake, to altitude of at least 4400 ft.*

* Present lowest part of basin floor is 3865 ft.altitude. Quaternary sedimentation has progressively raised the floor, and during late pre-Lake Lahontan time the floor probably was at least 200 to several hundred feet lower than now.

2. Subaerial sediments and a very strongly developed weathering profile of late pre-Lake Lahontan age (Paiute Formation and Cocoon Soil).

3. Deep-lake sediments (Eetza Formation), and minor intertonguing subaerial deposits, of early Lake Lahontan age.

4. Subaerial sediments, and intertonguing shallow-lake sediments (Wyemaha Formation) and a strong weathering profile (Churchill Soil), of mid-Lake Lahontan age.

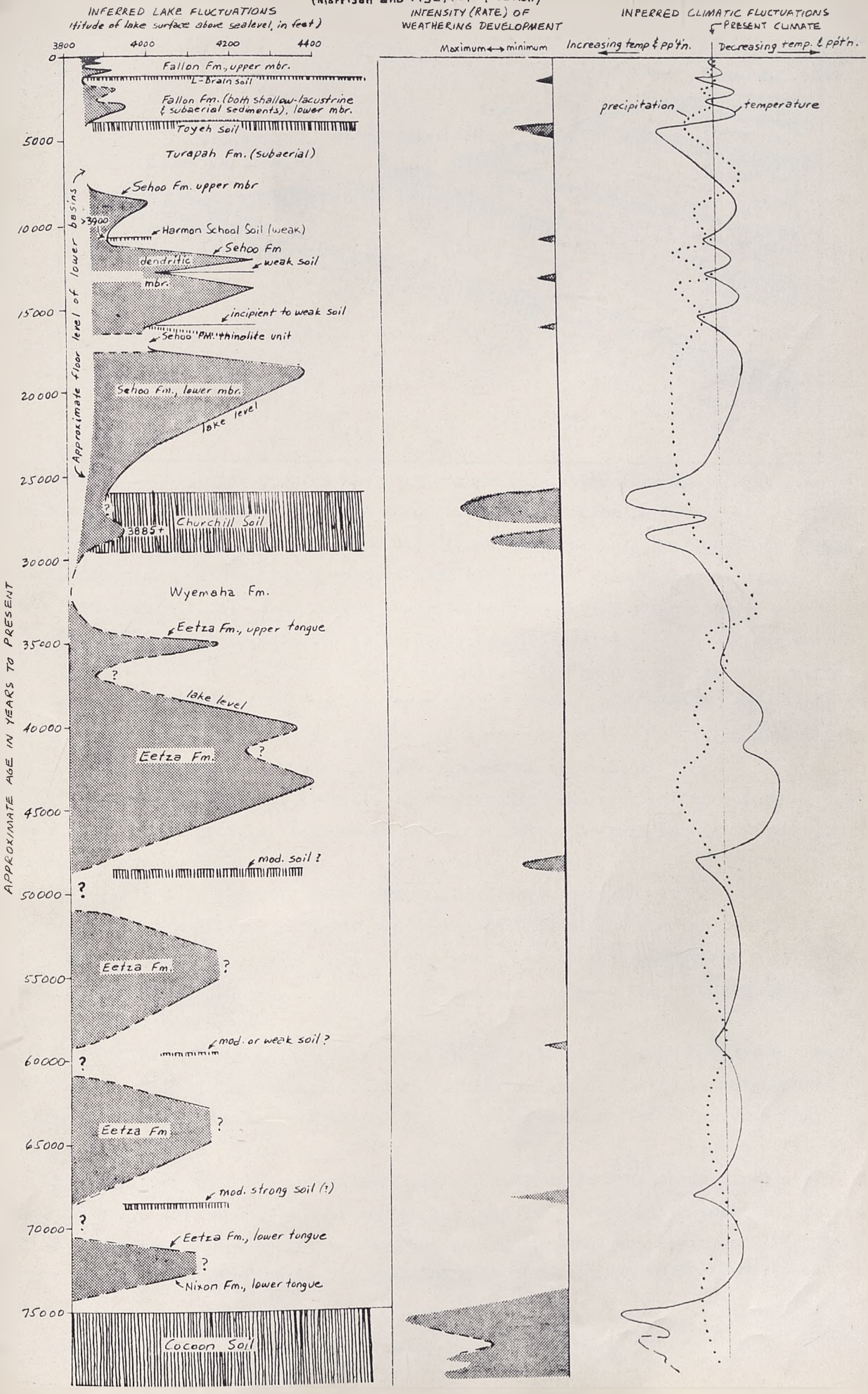
5. Deep-lake sediments (Sehoo Formation), and minor intertonguing subaerial deposits (Indian Lakes Formation), of late Lake Lahontan age.

6. Subaerial sediments (Turupah Formation)and a moderate weathering profile (Toyeh Soil), of early post-Lake Lahontan age.

7. Subaerial sediments and intertonguing shallow-lake sediments (Fallon Formation)of late post-Lake Lahontan age.

Of the seven weathering profiles that have been identified, five have been named formally and defined as soil-stratigraphic units, and qualify as geosols as this term is used in this paper. These are the Cocoon Soil (oldest), Churchill Soil, Harmon School Soil, Toyeh Soil, and L Drain Soil. Fig. 5.2 shows their general stratigraphic relations. Two other weakly developed soils are older than the Toyeh and Churchill Soils, respectively, and are known only from rare buried occurrences (intercalated with deposits of Sehoo and Eetza age), elsewhere, where relict, being masked by the stronger development of the younger soils. Descriptions of the profiles of each of the named soils, at their type localities in the southern Carson Desert area, near Fallon, Nevada, are given in the appendix. Chemical and mechanical analyses also are given for these profiles. Note that the concentrations of colloidal clay in the B horizons and of calcium carbonate in the Cca horizons increase markedly with increasing degree of development of the weathering profiles.

Fig. 5.1. Pluvial-lake oscillations compared with fluctuations in weathering rates and with changes in mean annual temperature and precipitation, inferred from the late Quaternary succession in the Lake Lahontan area, Nevada.
(Morrison and Frye, in preparation)



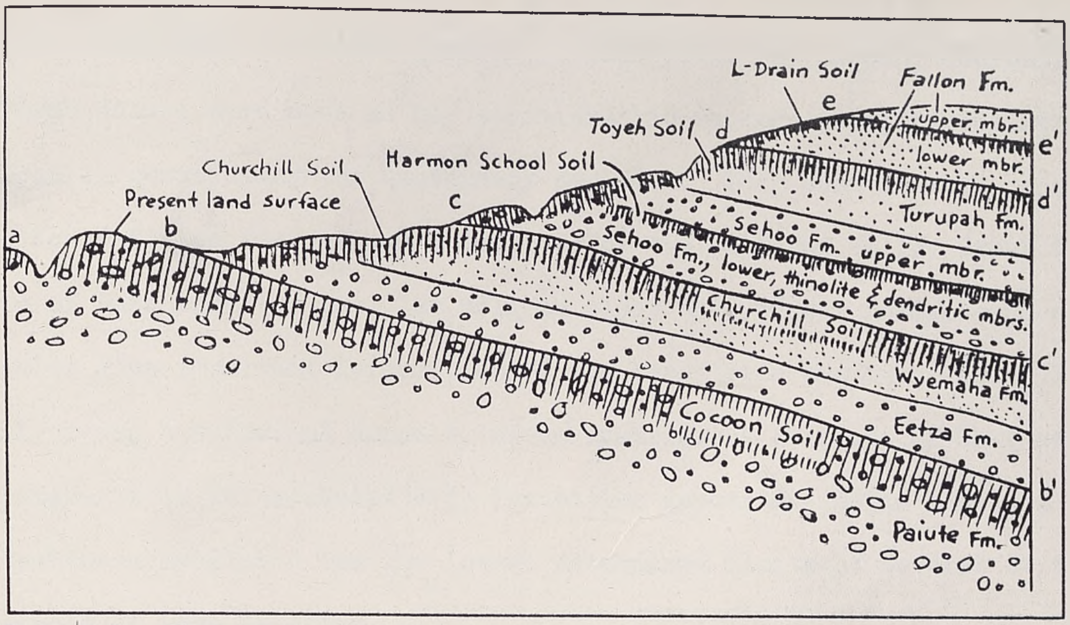


Fig. 5.2. Diagram showing the stratigraphic relations and age of the five main (named) late Quaternary soils (geosols) in the Carson Desert area.

The Cocoon soil (geosol) is the oldest and most strongly developed and occurs on deposits as young as the uppermost beds of alluvial gravel and colluvium of the Paiute formation. a-b, relict occurrence; b-b, buried occurrence.

The Churchill soil (geosol) is strongly developed, intermediate in degree of development between the Cocoon and Toyeh soils. It lies on deposits as young as the uppermost subaerial beds of the Wyemaha formation. b-c, relict occurrence; c-c, buried occurrence. It is not evident where it developed on relicts of the Cocoon soil because of the stronger development of the latter.

The Harmon School soil (geosol) is very weakly developed and therefore occurs only as a buried profile, because in relict occurrences it is masked by the more strongly developed Toyeh soil.

The Toyeh soil (geosol) is moderately developed compared with the other soils. It occurs on deposits as young as the uppermost beds of the Turupah formation. c-d, relict occurrence; d-d, buried occurrence. It is not evident where it developed on relicts of the Churchill and Cocoon soils owing to the stronger development of these soils.

The "L" Drain soil (geosol) is found on deposits as young as the second lake unit of the Fallon formation. It is so weakly developed that it is not evident where it developed on relicts of the older soils. d-e, relict occurrence; e-e, buried occurrence.

The deposits that are intermediate in age between each of these geosols are devoid of any evidence of coeval weathering profile development, except for several thin horizons of incipient to weak weathering (unnamed weathering profile) that occur locally as buried soils intercalated with the Eetzaand Sehoo formations.

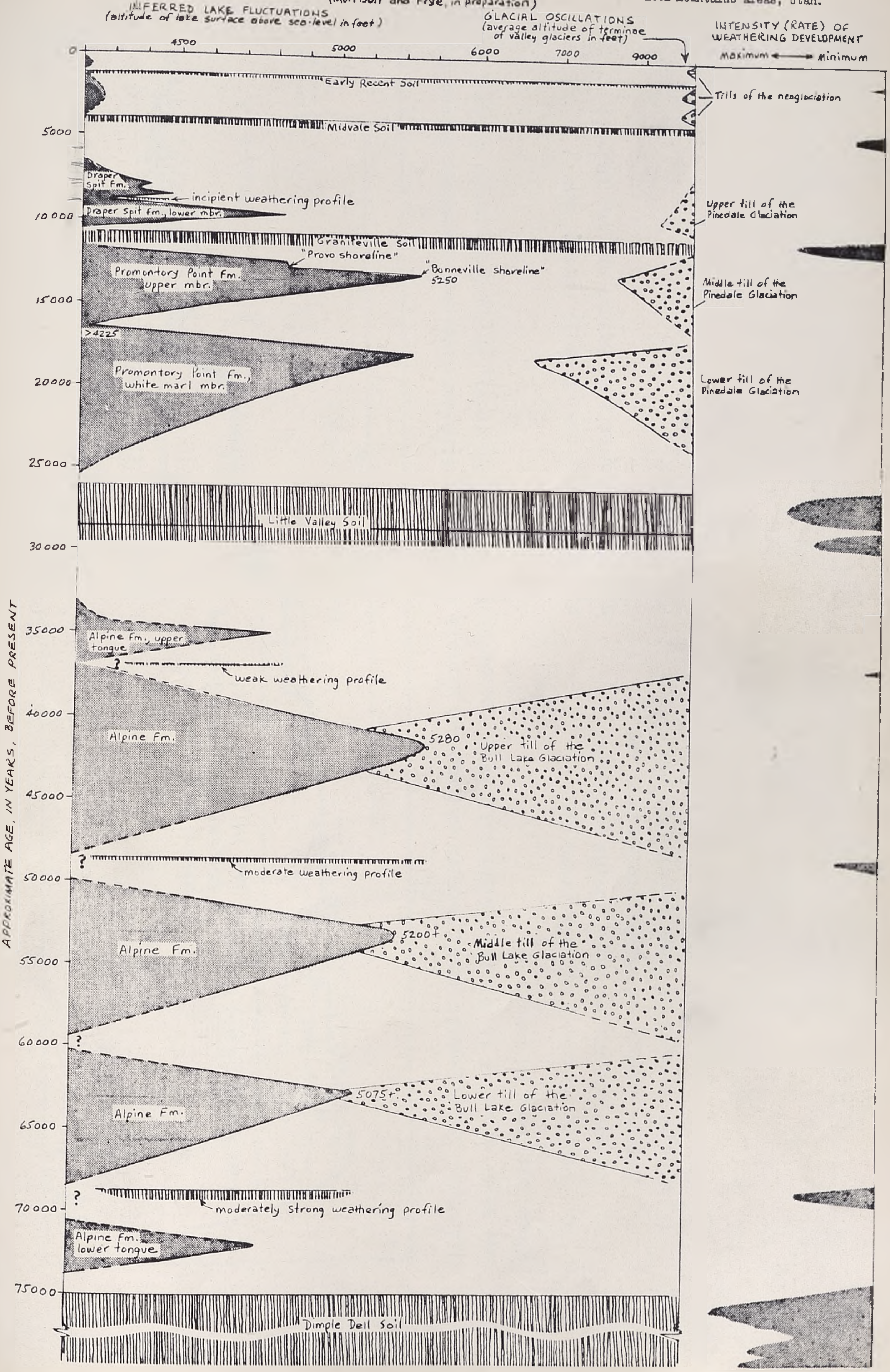
The three most prominent geosols in this area, the Cocoon, Churchill, and Toyeh Soils, were used as key stratigraphic markers in differentiating and locally correlating the Quaternary deposits. They constitute a major basis for defining the various formations of the Lahontan Valley Group (deposits of Lake Lahontan age), as well as older and younger formational units. They also give important information of the lacustrine history (Fig. 5.1). A soil, being a subaerial deposit, gives unambiguous evidence of lake recession where it is intercalated with lacustrine deposits. Soils supplied the best evidence available for the lowest determined limits of several of the recessions of Lake Lahontan.

The Toyeh Soil has been proposed as the logical marker for the boundary between the Pleistocene and Recent epochs (series) in the Great Basin region (Morrison, 1961d). This proposal is based on the fact that this soil and its correlatives (e.g., the Midvale Soil in the Lake Bonneville area) is the most widely identifiable and traceable stratigraphic marker unit, and also the most nearly time-parallel one, in the late Wisconsinan-Recent successions in all types of surficial terrains--glacial, pluvial lacustrine, alluvial, eolian, colluvial, etc.

Lake Bonneville area, Utah

The most recent stratigraphic studies of the deposits and history of Lake Bonneville, using weathering profiles as soil-stratigraphic units, are those in eastern Jordan Valley south of Salt Lake City, Utah, (which includes the famous area of glacial till and outwash intertonguing with Lake Bonneville sediments, below the mouth of Little Cottonwood Canyon)(Morrison, 1961a and in press, b), and also in Little Valley and vicinity, Promontory Point, Box Elder County, Utah (Morrison, in preparation). The latest classification of the deposits of Lake Bonneville and the late Quaternary soils is given in fig. 5.3 , which is based largely upon the above-mentioned two studies.

Fig. 5.3. Pluvial-lake and glacial oscillations compared with fluctuations in weathering rates, inferred from the late Quaternary successions in the Lake Bonneville and Wasatch Mountains areas, Utah. (Morrison and Frye, in preparation)



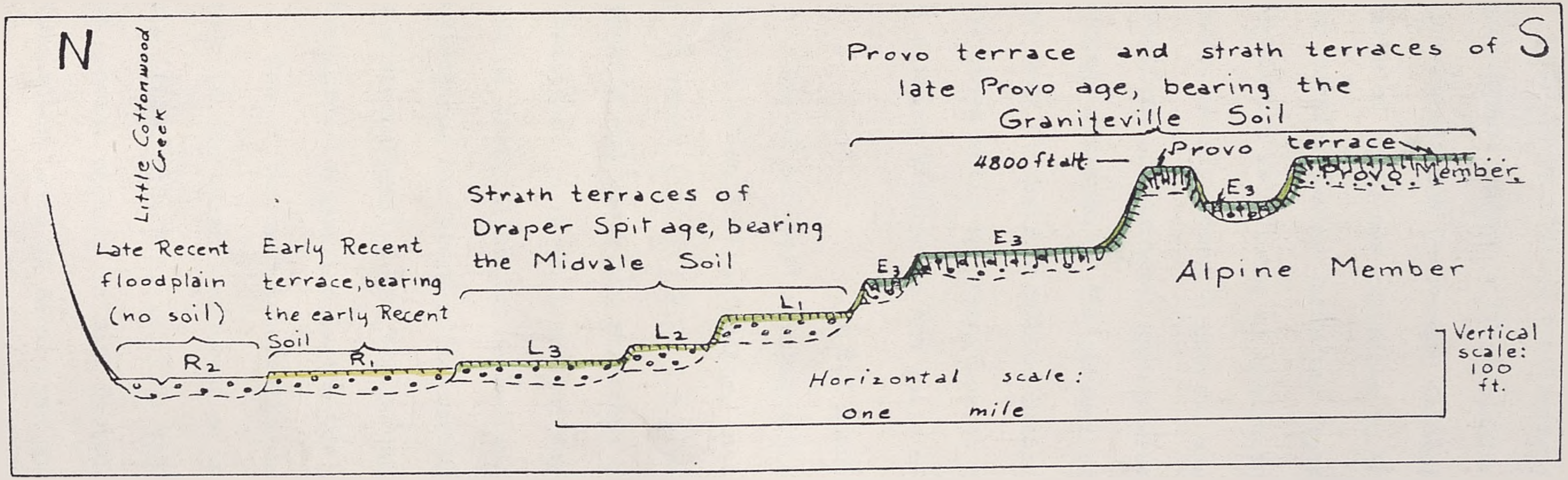


Fig. 5.4. Diagrammatic cross section showing how geosols are used to differentiate and correlate stream terraces and their deposits in the Lake Bonneville area (along Little Cottonwood Creek, in eastern Jordan Valley, Utah). (From Morrison, in press, b.)

E_3 Strath terraces of late Provo age

L_1 Strath terrace of early Draper Spit age

L_2 Strath terrace of middle Draper Spit age

L_3 Strath terrace of late Draper Spit age

R_1 Strath terrace of early Recent age

R_2 Floodplain of late Recent age (no geosol)

In these studies weathering profiles (geosols) were used for local rock-stratigraphic differentiation and correlation (figs. 5.3 and 5.4), and the various Quaternary formational units were defined largely on the basis of their age relations to the soils. Each of the various soils has several facies variants, depending upon local precipitation-altitude-vegetation relations; at altitudes below 4400-6000 ft. Pedocals predominate, mainly Brown, Sierozem, and Desert Soils; at higher altitudes Pedalfers predominate, mainly Brown Podzolic Soils and Planosols. The average altitude of the pedocal-pedalfer facies change is highest for the very strongly developed Dimple Dell Soil and successively lower for the other soils in proportion to their relative degree of development, in similar fashion to the soils of the La Sal Mountains area (see below).

La Sal Mountains area, Utah

One of the best stratigraphic studies of Quaternary deposits, involving use of soil stratigraphy, is that of the La Sal Mountains area, Utah, by Richmond (1962). Fig. 5.5 gives the stratigraphic column of Quaternary deposits and shows the relations of the geosols (soils) to the various formations. A given formation may comprise various lithogenetic facies such as till, alluvium, colluvium, or eolian sediments, with mutually intertonguing or intergrading relations. The base of a formation is commonly a disconformity; its top is marked by a distinctive soil whose highest stratigraphic position is the top of this formation.

One of the most interesting and valuable contributions from Richmond's study is the careful observation of the facies relations of the various geosols (figs. 4.8, 5.6, and 5.7; tables 5.2, 5.3, 5.4, and 5.5)--not merely the general changes in soil facies with altitude, but also in specific characteristics, such as color (hue and chroma), maximum thickness of B or Bca

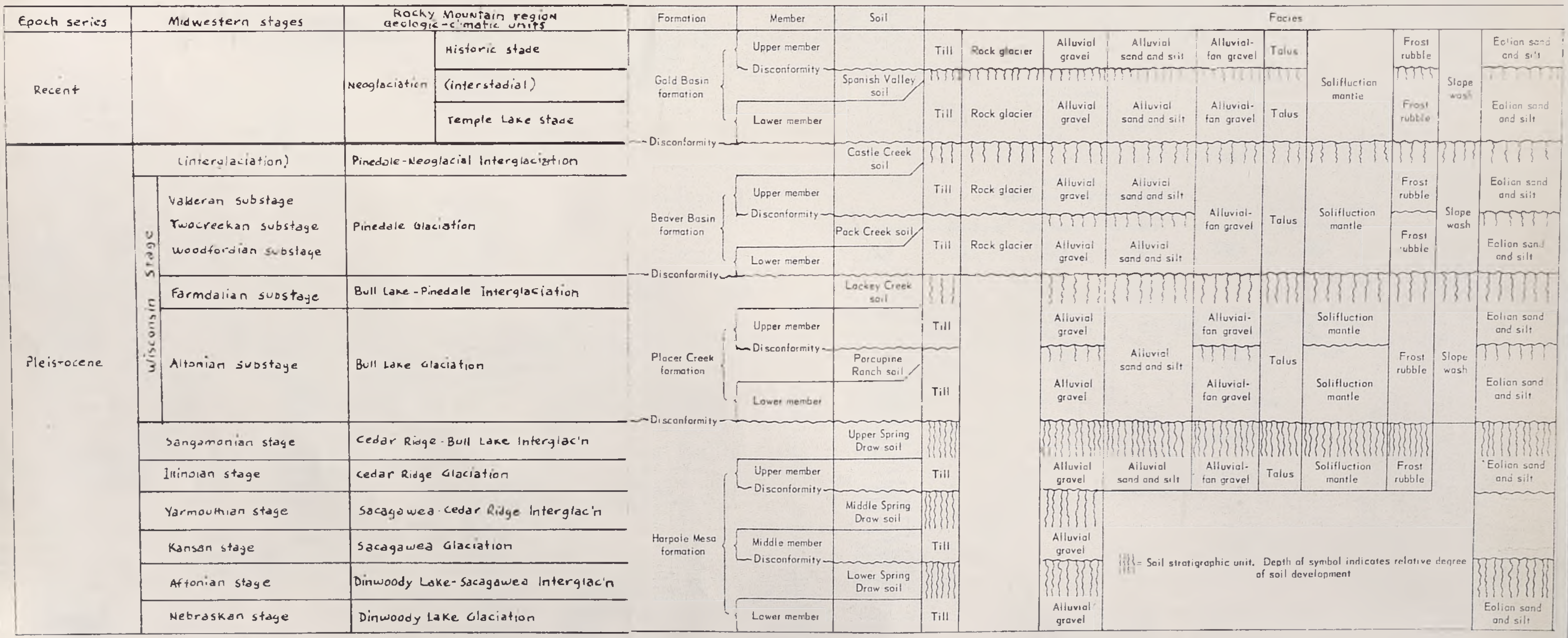


Fig. 5.5. Stratigraphic succession of Quaternary deposits in the La Sal Mountains, Utah. (Modified after Richmond, 1962a.)

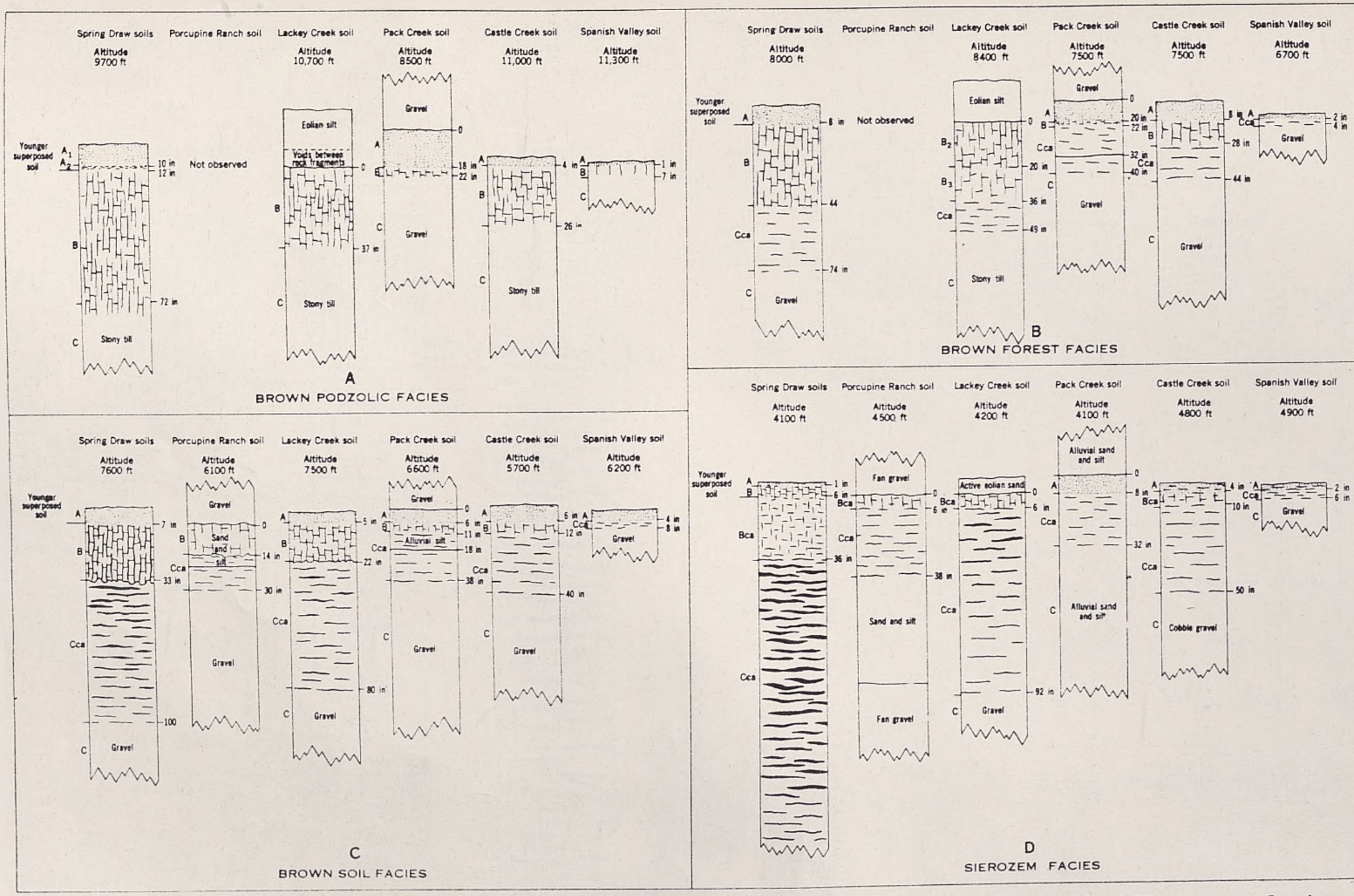
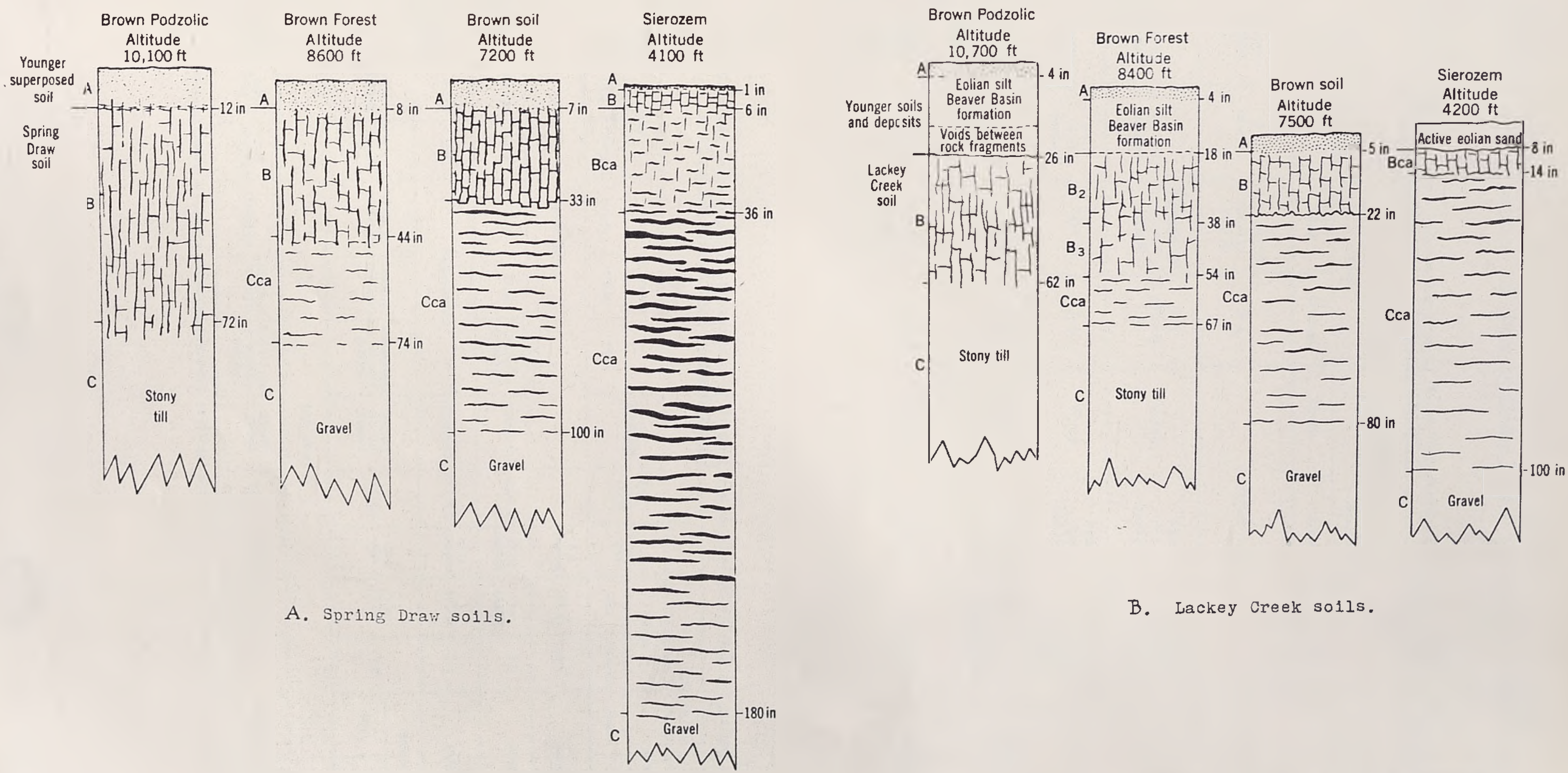


Fig. 5.6. Comparison of the general character and thickness of facies of soils of different geologic age in the La Sal Mountains, Utah. A, Brown Podzolic facies; B, Brown Forest facies; C, Brown soil facies; D, Sierozem facies. The Spring Draw soils are very strongly developed and of pre-Wisconsin age; the Porcupine Ranch is moderately developed and of intra-Bull Lake (intra-Early Wisconsin) age; the Lackey Creek soil is strongly developed and of inter-Bull Lake-Pinedale (mid-Wisconsin) age; the Pack Creek soil is weak and of intra-Pinedale (Brady?) age; the Castle Creek soil is moderately developed and of altithermal age; and the Spanish Valley soil is weak and of early Recent age. (From Richmond, 1962a.)



A. Spring Draw soils.

B. Lackey Creek soils.

Fig. 5.7. Diagrammatic, typical profiles of the Spring Draw and Lackey Creek soils (which are of pre-Wisconsinan and mid-Wisconsinan age, respectively), showing their general character and thickness in four major facies (from Richmond, 1962a).

horizon, maximum thickness of Cca horizon, and reaction (pH) (fig. 4.9).

TABLE 5-2.—*Comparison of profile characteristics of Brown Podzolic facies of soils of different geologic age* in the La Sal Mts., Utah (Richmond, 1962a)

Soil	B horizon								
	Altitude range (feet)	Thickness	Color		Structure			Plasticity	Reaction (pH)
			Hue	Chroma	Grade	Class	Type		
Spanish Valley ("Early Recent")	Above 7,000	4–6 inches	7.5–10*YR*, mostly 10*YR*	2–4	Structureless			Nonplastic	6.0–7.0
Castle Creek ("Altithermal")	Above 8,000	10–22 inches	7.5–10*YR*, mostly 10*YR*	2–4, mostly 4	Weak	Fine to medium	Crumb to subangular blocky	Nonplastic	5.0–6.5
Pack Creek ("Brady")	Above 7,800	4–8 inches	10*YR*	2–4,	Weak	Fine to medium	Subangular to angular blocky	Nonplastic	6.0–6.5
Lackey Creek ("Mid wisconsin") ("Post-Bull Lake")	Above 8,400 to 9,000	3.0–4.5 feet	5–10*YR*, mostly 7.5*YR*	2–6, mostly 6	Weak to moderate	Fine to medium	Angular to subangular	Mostly slightly plastic	4.5–6.5
Spring Draw ("Pre-Wisconsin")	Above 8,600 to 9,200	4.5–8.0 feet	10*R*–7.5*YR*, mostly 5*YR*	2–6, mostly 6	Moderate to strong	Coarse to medium	Angular blocky	Plastic	4.0–5.5

TABLE 5.3—*Comparison of profile characteristics of Brown Forest facies of soils of different geologic age* in the La Sal Mts., Utah (Richmond, 1962 a)

Soil	Altitude range (feet)	B horizon								Cca horizon						
		Thickness	Color		Structure			Plasticity	Reaction (pH)	Thickness	Color		Structure			Reaction (pH)
			Hue	Chroma	Grade	Class	Type				Hue	Chroma	Grade	Class	Type	
Spanish Valley[1] ("Early Recent")	7,000–6,500	No B horizon								2–6 inches	10*YR*	3–4	Structureless			about 7.0
Castle Creek ("Altithermal")	7,800–7,000	4–10 inches	7.5*YR*	3–4	Weak	Fine	Granular to blocky	Nonplastic	6.5–7.3	12–24 inches	7.5–10*YR*	3–4	Structureless			7.0–7.5
Pack Creek ("Brady")	8,000–7,200	2–4 inches	10*YR*	3–4	Structureless			Nonplastic	6.1–6.5	6–12 inches	10*YR*	3–4	Structureless			7.0–7.5
Lackey Creek ("Mid-Wisconsin") ("Post-Bull Lake")	9,000–7,400	2.5–3 feet	2.5–7.5*YR*	3–6, mostly 4.5	Weak to moderate	Fine to medium	Subangular to angular blocky	Mostly slightly plastic	6.1–7.0	1.5–3 feet	7.5–10*YR*	3–6	Structureless to— Weak	Fine to medium	Platy	7.0–8.0
Porcupine Ranch ("Intra Bull-Lake")	Not observed															
Spring Draw ("Pre-Wisconsin")	9,200–8,200	4–5 feet	10*R* 2.5*YR*	2–6 mostly 6	Moderate	Medium to coarse	Subangular to angular blocky	Mostly plastic	5.5–6.2	3–4 feet	2.5–7.5*YR*	3–6	Weak to moderate	Medium to coarse	Blocky to platy	7.0–8.0

[1] Spanish Valley soil has an azonal alkaline A–C profile, whose characteristics in an arbitrary altitude range are given here under Cca horizon.

TABLE 5.4—*Comparison of profile characteristics of Brown soil facies of soils of different geologic age* in the La Sal Mts., Utah (Richmond, 1962a).

Soil	Altitude range (feet)	B horizon: Thickness (inches)	B horizon: Color: Hue	B horizon: Color: Chroma	B horizon: Structure: Grade	B horizon: Structure: Class	B horizon: Structure: Type	B horizon: Plasticity	B horizon: pH	Cca horizon: Thickness (inches)	Cca horizon: Color: Hue	Cca horizon: Color: Chroma	Cca horizon: Structure: Grade	Cca horizon: Structure: Class	Cca horizon: Structure: Type	Cca horizon: pH
Spanish Valley ("Early Recent")	[1] 6,500 5,500	No B horizon								2–6	5–10*YR*	3–4	Structureless			About 7.5
Castle Creek ("Altithermal")	7,200–5,400	6–12	7.5–10*YR*, mostly 10*YR*	3–4	Weak	Fine	Columnar to blocky	Nonplastic	7.0–7.8	24–36	7.5–10*YR*	3–4	Structureless or— Weak to strong	Fine	Blocky to columnar	8.0–8.5
Pack Creek ("Brady")	7,000–5,100	4–8	5–10*YR*	3–4	Structureless or— Weak	Fine	Columnar to blocky	Nonplastic to slightly plastic	7.0–7.5	12–24	5–10*YR*	3–4	Structureless			7.5–8.5
Lackey Creek ("Mid-Wisconsin") ("post-Bull Lake")	7,800–5,500	24–36	5–10*YR*, mostly 5*YR*	4	Weak to moderate	Fine to coarse	Columnar to blocky, mostly blocky	Nonplastic to slightly plastic	6.8–7.5	36–60	5–10*YR*	2–4	Structureless or— Weak to moderate	Strong	Platy	8.0–9.0
Porcupine Ranch ("Intra Bull-Lake")	[2] 6,800 5,400	6–24	5–7.5*YR*	3–4	Weak	Fine to medium	Angular to subangular blocky	Nonplastic to plastic	6.5–7.2	12–30	7.5*YR*	4	Structureless			7.5–8.0
Spring Draw (Soils) ("Pre-Wisconsin")	8,500–5,800	40–50	2.5–10*YR*	4–6, mostly 4	Moderate to strong	Medium to coarse	Columnar to angular blocky	Plastic to very plastic	6.0–7.0	48–72	5–7.5*YR*	1–4	Coarse	Massive or medium to strong	Blocky to platy	8.0–9.0

[1] Spanish Valley soil has an azonal alkaline A-C profile, whose characteristics in an arbitrary altitude range are given here under Cca horizon.
[2] Range in which soil is preserved; probably not equivalent to original extent.

TABLE 5.5—*Comparison of profile characteristics of Sierozem facies of soils of different geologic age* in the La Sal Mts., Utah (Richmond, 1962a)

Soil	Altitude range (feet)	B or Bca horizon: Thickness	B or Bca horizon: Color: Hue	B or Bca horizon: Color: Chroma	B or Bca horizon: Structure: Grade	B or Bca horizon: Structure: Class	B or Bca horizon: Structure: Type	B or Bca horizon: Plasticity	B or Bca horizon: Reaction (pH)	Cca horizon: Thickness	Cca horizon: Color: Hue	Cca horizon: Color: Chroma	Cca horizon: Structure: Grade	Cca horizon: Structure: Class	Cca horizon: Structure: Type	Cca horizon: Reaction (pH)	Cca horizon: Cementation
Spanish Valley ("Early Recent")	below 5,500 [1]	No B or Bca horizon								2–6 inches	5–7.5*YR*	3–4	Structureless			7.5–8.0	None.
Castle Creek ("Altithermal")	below 5,400	About 6 inches	5–10*YR*	4	Structureless or weak platy			Nonplastic	8.0–8.5	36–48 inches	5–7.5*YR*	3–4	Structureless			8.5–9.0	None.
Pack Creek ("Brady")	below 5,100	No B or Bca horizon								About 24 inches	5–10*YR*	3–4	Structureless			8.0–8.5	None.
Lackey Creek ("Mid-Wisconsin") (post Bull-Lake")	below 5,500 to 5,800	6–24 inches	2.5–7.5*YR*	4	Structureless or— Moderate	Medium	Blocky	Slightly plastic	9.0	65–85 inches	2.5–7.5*YR*	2–4	Structureless or— Medium to fine	Strong	Platy	8.8–9.0	Weak.
Porcupine Ranch ("Intra Bull-Lake")	[2] 4,700–4,200	6± inches	2.5*YR*	6	Structureless			Nonplastic	8.0–8.5	24–36 inches	2.5*YR*	4	Structureless or— Weak	Coarse	Columnar	8.0–8.5	None.
Spring Draw ("Pre-Wisconsin")	below 5,800 to 6,200	24–36 inches	5*YR*	3	Structureless			Plastic	9.0	10–18 feet	2.5–7.5*YR*	1–4	Structureless or— Strong	Coarse	Platy	9.0	Weak to strong.

[1] Spanish Valley soil has an azonal alkaline A-C profile, whose characteristics in an arbitrary altitude range are given here under Cca-horizon.
[2] Range in which soil is preserved, probably not equivalent to original extent.

CHAPTER VI

APPLICATION OF SOIL STRATIGRAPHY TO LONG-DISTANCE TIME-STRATIGRAPHIC CORRELATION OF QUATERNARY DEPOSITS

Special problems in correlating Quaternary deposits: Quaternary correlation is different!

In the past it has proved very difficult to correlate the finer subdivisions of Quaternary deposits consistently and reliably over long distances, particularly between varied lithogenetic terrains such as continental glacial, mountain glacial, alluvial terrace, marine shore-terrace, and pluvial lake landscapes. Paleontology, usually the chief aid in such problems in older parts of the geologic column, is of little assistance, both on account of the scantiness of faunal and floral remains in many deposits, and, more importantly, because the finer time-stratigraphic subdivisions (stage and smaller) of the Quaternary span time intervals that were too short in duration for appreciable evolution of species to have taken place. Thus, the fossil record, even where present, is not diagnostic for fine age placement and correlation. Some fossil records, particularly molluscan and pollen ones, locally have been used successfully for fine subdivision of Quaternary sequences, but purely on an ecologic basis--i.e., the biozones are controlled primarily by environment instead of evolution. Used on this basis the fossil record affords merely rock-stratigraphic or geologic-climatic subdivision, and such units are obviously time-transgressive.

Correlation by means of geologic-climatic units

The traditional means of long-distance correlation of Quaternary deposits is based on matching sequences of depositional cycles, or sequences of cycles of climatic change that can be inferred from the depositional record. The stratigraphic units that express these periodic climatic-depositional fluctuations are now termed geologic-climatic units (Amer. Comm. on Stratigraphic Nomenclature, 1961, p. 660).

Climatic change was probably the outstanding characteristic of Quaternary time (in contrast to most of earlier geologic time). Periodic climatic fluctuations of varying amplitude and duration caused changes in rate and type of various surficial erosional and depositional processes and these variations are identifiable in the geologic record. The basic climatic oscillations were worldwide and probably essentially synchronous, although their local manifestations were influenced by latitude, altitude, continentality, storm tracks, etc. The effects of the climatic oscillations upon the depositional-erosional record are most marked at temperate and high latitudes (and at high altitudes at low latitudes) and has resulted in the Quaternary depositional record being in a sense unique and unlike most of the rest of the geologic column. The climatic-depositional cycles are manifest by fluctuations in surficial processes such as advances and retreats of glaciers, rises and falls of pluvial lakes, eustatic shifts in sea level, aggradation and degradation along streams, soil development, etc. Generally some idea can be gained of the amplitude and duration of the various depositional cycles from the distribution, thickness and lithologic character of the deposits that record them. By matching the successions of these cycles in separate areas, the sequences of deposits that record them can be correlated on a geologic-climatic basis.

Various uncertainties, however, confound attempts at long-distance correlation solely by this means. Correlation may be hampered by the complexity of the record in one area and its meagreness in another; by portions of the record being missing due to accidents of erosion, nondeposition, or lack of exposure; and by distortion of the record by such non-climatic variables as diastrophism. Commonly in such cases one cannot be certain that he has correctly identified the same depositional cycle (or geologic-climatic unit) in the sequences in different areas.

The main defects of the geologic-climatic means of correlation lie in the inherent non-time-parallelism of geologic-climatic units (Amer. Comm. on Stratigraphic Nomenclature, 1961, p. 660), and in the fact that the various types of depositional cycles that may result from a given climatic cycle may be of different duration and also may be out of phase with each other. For example: till of the Woodfordian stade (substage) of the Wisconsinan Glaciation (Stage) spans many more thousands of years in southern Ontario than it does at the Shelbyville-White Rock end moraines in central Illinois. Pluvial lake cycles cannot be assumed to be exactly synchronous and of the same duration as glacial cycles, even in the same region. In extreme cases, two sets of climatic-depositional cycles (geologic-climatic units) may be as much as 100% out of phase with each other--for example, cycles of glaciation compared with eustatic shifts in sea level. In the Rocky Mountains and other mountain systems of the western U. S., strong alluviation and stream aggradation took place during the various glacial maxima; on the Mississippi delta and at sea-mouths of other rivers, on the other hand, marked down-cutting occurred during the glacial maxima because of eustatic drop in sea level.

Correlation by using geosols as time-stratigraphic markers

Many of the defects inherent in geologic-climatic correlation can be minimized by using geosols as time-stratigraphic markers. Geosols are not only excellent rock-stratigraphic markers for differentiation and correlation in local areas, but also they are commonly the best, and certainly the most ubiquitous, time-stratigraphic markers for long-distance correlation of Quaternary successions in all types of lithogenetic terrains.

Validity of geosols as time-stratigraphic markers

The validity of geosols as time-stratigraphic markers for long-distance correlation hinges upon whether they can be proved to be reasonably time-parallel. The first line of proof is based on comparison of the records of Quaternary successions (rock-stratigraphic units and geosols) in the numerous areas throughout temperate North America where these successions have been comprehensively studied. (The examples given in chapter 5 are typical.) The stratigraphic positions of geosols of similar relative development in relation to the rest of the stratigraphic successions match so well between the various areas as to strongly suggest that the geosols of similar relative development are at least widely traceable rock-stratigraphic markers. Moreover, in every case the physical record shows that each geosol has a separate identity, and that the instability deposits intervening between successive geosols evince no weathering-profile development, although in many cases their intervals of deposition obviously lasted longer than the weathering optima. Thus, the weathering optima must have been comparatively brief and widely spaced, and therefore occupied only very small parts of total Quaternary time. This clearly means that the rate of pedogenic weathering varied greatly at these latitudes, from negligible for long intervals to several orders of magnitude greater during the stronger pedogenic intervals. Evidently, therefore, the geosols formed in response to relatively infrequent combinations of climatic factors that induced erosional stability (probably mainly because of luxuriant plant cover) and a more rapid rate of chemical weathering than normal.

In conclusion, the stratigraphic records in many areas testify that the weathering optima were periodically repeated parts of whole climatic cycles--mainly fluctuations of temperature and precipitation--that are manifest in Quaternary sequences of each area. The most strongly developed geosols formed during the later parts of the main inter-glacial and pluvial lake-desiccation intervals; weaker geosols formed during some of the shorter and warmer recession intervals; and weathering at other times was inappreciable.

The principal factor causing the accelerated weathering during any weathering optimum may have been increased temperature (see chapter 7)--appreciably higher than temperatures today, particularly for the main weathering optima. Climatologists recognize that secular temperature variations are world-wide and synchronous, inasmuch as they inherently result from changes in solar energy, whereas changes in precipitation tend to be much less ubiquitous and synchronous (Brooks, 1949; Willett, 1949, 1951, 1953, p. 63; Flohn, 1950, 1952; Wexler, 1953, Wolbach, 1953; Bell, 1953; Butzer, 1957a, 1957b, 1961, p. 36-37, 44, 48; Schwarzbach, 1961, p. 229-230). This fact and the comparative brevity of the weathering optima strongly suggests that these intervals were essentially synchronous throughout temperate latitudes; in other words, that geosols are time-parallel.

Further testimony that the weathering optima were indeed brief and also essentially synchronous over wide regions comes from comparing the radiocarbon chronologies of various areas where the duration of some of the main weathering optima can be closely bracketed. The fullest information to date is from the upper Mississippi Valley and Lake Lahontan areas, and is given in another paper (Morrison and Frye, in preparation). It appears to establish beyond reasonable doubt that the weathering optima when the "altithermal" and middle Wisconsinan (Farmdalian) geosols formed were less than 1,000 and 4,000 years long, respectively, and essentially synchronous between the two areas.

Probably slight non-time parallelism occurs in geosols due to variations in latitude and altitude (and possibly other local climatic influences), but these differences appear to be relatively small, again because of the synchronous, world-wide character of the comparatively brief maxima of solar energy that induced the pedogenic intervals. There is no evidence, stratigraphic or otherwise, that geosols are time-transgressive to anywhere near the same degree as many Quaternary rock-stratigraphic and geologic climatic units. Thus, geosols in temperate latitudes are believed to be more closely time-parallel than any other widespread type of stratigraphic unit in Quaternary deposits; much more so than the glacial, lacustrine, or fluvial deposits with which they are associated.

For the above reasons, I disagree with the majority opinion of the Pleistocene subcommittee of the American Commission on Stratigraphic Nomenclature (Richmond, 1959, p. 669). This subcommittee concluded that soils (weathering profiles) may be no more time-parallel than their glacial and pluvial conterparts, and should not define time-stratigraphic (stage) or geologic-time (age) units. On the contrary, I propose that geosols are commonly the best time-stratigraphic markers and that they are valid not just within small regions, but also inter-regionally, for long-distance correlation throughout temperate latitudes. Furthermore and as a corollary, in many cases geosols are useful and valid for defining the boundaries of time-stratigraphic units (stages and substages) within the Quaternary.

In view of their distinctive, widely traceable characteristics and utility as time-stratigraphic markers some geosols have achieved recognition on a regional basis. This is true, for example, of the geosol that is known informally in the Great Basin and Rocky Mountain regions as the "altithermal soil"[1/] [it still lacks a generally recognized formal name, although locally it has been formally named, as in the Lake Lahontan, Lake Bonnevile and La Sal Mountain areas (see chapter 5)]. Other examples are the Sangamon, Yarmouth, Afton Soils, and even the Brady Soil, in the Midwest and Great Plains.

1/ See p. 129 for definition.

Technique of long-distance correlation by use of geosols

In making long-distance time-stratigraphic correlation by means of geosols, the total Quaternary sequence (rock-stratigraphic units and geosols) of each area should be known as fully as possible, particularly noting the relative position (age) and relative degree of development of the geosols in each area. The basic framework for correlation is provided by matching the stronger geosols of each area as closely as possible in terms of their relative age and relative degree of development. The stronger geosols are generally the most readily recognized in the successions of each area, and also are the most readily evaluated in terms of their relative development within a given succession of geosols.

After the main geosols are correlated with each other, the rock-stratigraphic (or geologic-climatic) units and weaker soils that are intermediate in age between the main geosols of this framework are then correlated by matching those units that record depositional cycles (or parts of cycles) of similar relative age, magnitude, and climatic genesis. Thus, units recording early lake cycles are correlated with each other and with units recording early glacial cycles; likewise, lake-recessional units are correlated with glacial-recessional units. It should be recognized, of course, that such correlations are influenced by the completeness of detail and accuracy of the stratigraphic records of the areas being correlated. Weakly developed geosols are less useful than strong ones as time-stratigraphic markers because they are less easily identified, and also older weak geosols occur (unmodified by later intervals of stronger weathering) only in buried occurrences.

Summarizing the technique of correlation by soil stratigraphy: first, the stronger geosols of similar relative development and relative age are correlated; then, between the main geosols, the deposits that record depositional cycles of similar age, climatic genesis, and magnitude are correlated, as are any weakly developed geosols.

Correlation of Quaternary deposits of the upper Mississippi Valley, Great Plains, Rocky Mountains, Lake Bonneville, and Lake Lahontan areas, using geosols as time-stratigraphic markers

Several regional and inter-regional correlations of Quaternary sequences have been published that use soils (=geosols) as inter-regional time-stratigraphic markers. These include a correlation between the deposits of the La Sal Mountains, Utah, and the Lake Bonneville and Lake Lahontan areas (Richmond, Morrison, and Bissell, 1952); of alluvial deposits from New Mexico and Arizona to Wyoming in the Rocky Mountain region (Leopold and Miller, 1954); of deposits in the La Sal Mountains, Utah, with those in other areas of mountain glaciation in Colorado, in the Wind River Mountains, Wyoming, and in Nebraska and Kansas (Richmond, 1962a); of surficial deposits in the vicinity of Denver, Colorado, with those in Nebraska (Scott, 1963); and the pluvial lacustrine successions of Lakes Lahontan and Bonneville with the glacial sequence of the Sierra Nevada (Morrison, in press, a). Most of these correlations have used several soils (geosols) as time-stratigraphic markers, which improves their reliability. (The correlation by Leopold and Miller probably is least reliable because it is based largely upon just one soil, the weathering profile of "altithermal" age.) In the correlations that have used several geosols as the basic framework for correlation, any mistakes in correlation (some already are apparent) have resulted from incomplete knowledge of the stratigraphic successions in the local areas, rather than from defects in the basic principles of soil-stratigraphic correlation.

Currently, a new, unusually detailed correlation is being made of the middle and late Quaternary deposits of the upper Mississippi Valley, Great Plains, Rocky Mountains, Lake Bonneville, and Lake Lahontan areas (Morrison and Frye, in preparation). A new correlation is appropriate because considerable new knowledge is available on the Quaternary stratigraphy of all these areas, that necessitates revision of their stratigraphic successions. Details of this correlation will be published elsewhere, but are summarized below and in table 6.2.

According to this correlation scheme, the Cocoon Soil of the Lake Lahontan area is correlated with the Dimple Dell (younger pre-Lake Bonneville) Soil of the Lake Bonneville and Wasatch Mountain areas, and with the Sangamon Soil in Illinois. The Churchill Soil of the Lake Lahontan area is correlated with the Little Valley (post-Alpine and post-Bull Lake) Soil of the Lake Bonneville and Wasatch Mountain area, and with an unnamed weathering profile of Farmdalian age in Illinois. The Toyeh Soil is correlated with the Midvale (post-Pinedale) Soil of the Lake Bonneville--Wasatch areas and with an unnamed "Recent" weathering profile in Illinois.

The deposits intermediate in age between these main geosols are correlated as follows: The Eetza Formation of Lake Lahontan is correlated with the Alpine Formation of Lake Bonneville, with drift of the Bull Lake Glaciation in the Wasatch Mountains, and with the Altonian drift in Illinois. The Sehoo Formation of Lake Lahontan likewise is correlated with the Promontory Point and Draper Spit Formations of Lake Bonneville, with drift of the Pinedale Glaciation in the Wasatch Mountains, and with the Woodfordian and Valderan drifts in Illinois and Ontario. Finer details of the correlations can be seen in table 6.2.

TABLE 6.1—*Correlation of Quaternary deposits of the La Sal Mountains*
(Richmond, 1962a)

Midcontinent region time-stratigraphic standard			La Sal Mountains, Utah		San Juan Mountains, Colo		Front Range, Cache La Poudre River, Colo	Wind River Mountains, Wyo		Hopi Country, Arizona	Nebraska	Kansas	
			This report		Modified after Atwood and Mather (1932), Richmond (1954)		Modified after Bryan and Ray (1940), Ray (1940)	Modified after Blackwelder (1915), Richmond (1948), Moss (1949, 1951)		Modified after Hack (1942)	Modified after Schultz, Loeninghoener, and Frankforter (1951)	After Frye and Leonard (1952)	
Recent series			Gold Basin formation	Upper member		Rock glaciers	Modern moraines R2		Modern moraines	Naho formation	Flood plain deposits, Soil Z' alluvium		Flood plain and Recent low terraces
				Spanish Valley soil		Very weak soil	Very weak soil		Very weak soil	Very weak soil	Soil Z		
				Lower member		Moraines or rock glaciers	Type Sprague moraine R1		Temple Lake stage	Tsegi formation	Alluvium		
				Castle Creek soil		Modern soil	Moderate soil		Moderate soil	Moderate soil	Soil Y		Post-Bignell soil
Pleistocene series	Wisconsin stage	Late Wisconsin	Beaver Basin formation	Upper member	Wisconsin stage	Upper till	Type Long Draw deposit Moraine at Long Draw Reservoir W4	Pinedale stage	Upper till	Jeddito formation	Bignell	Sanborn formation	Bignell
				Pack Creek soil							Soil YY		Leached zone
				Lower member		Lower till	Type Corral Creek moraine W3		Lower till		loess		member
		Middle Wisconsin		Lackey Creek soil	Durango stage	Strong soil	Strong soil	Bull Lake stage	Strong soil		Brady soil (soil X)		Brady soil
		Early Wisconsin	Placer Creek formation	Upper member		Upper till	Home moraine W2		Upper till		Peorian		Peoria
				Porcupine Ranch soil							Soil W		Leached zone
				Lower member		Lower till	Pre-Home deposits W1		Lower till		loess		member
	Sangamon (?) stage		Harpole Mesa formation	Upper Spring Draw soil				Buffalo stage	Very strong soil		Sangamon soil		Sangamon soil
	Illinoian (?) stage			Upper member					Post-canyon till (?)		Loveland loess		Loveland member
	Yarmouth (?) stage			Middle Spring Draw soil		Very strong soil	Very strong soil				Yarmouth soil		Yarmouth soil
	Kansan (?) stage			Middle member	Cerro stage	Till	Prairie Divide till		Pre-canyon till (?)		Kansan till		Kansan till
	Aftonian (?) stage			Lower Spring Draw soil							Aftonian soil		Aftonian soil
	Nebraskan (?) stage			Lower member					Pre-canyon till (?)		Nebraskan till		Nebraskan till

Table 6.2. Correlation chart (Morrison and Frye, in preparation).

Approximate age in C^{14} years before present	MIDCONTINENT REGION TIME-STRATIGRAPHIC STANDARD	EASTERN MIDWEST Illinois - Wisconsin - Ontario Physical stratigraphic units	SOUTHWESTERN GREAT PLAINS West Kansas - West Texas Physical stratigraphic units	CENTRAL ROCKY MOUNTAINS Colorado - Utah Geologic-climatic and soil-stratigraphic units	LAKE BONNEVILLE AREA, UTAH Physical stratigraphic units	LAKE LAHONTAN AREA, NEV. Physical stratigraphic units
	Recent Series	alluvium and eolian sand	recent soil alluvium, eolian sand, basin deposits	NEOGLACIATION: Historic stade post-Temple Lake soil Temple Lake stade	Early Recent soil alluvium, colluvium, eolian sand, valley-fill sediments	Fallon Fm.: upper member L Drain soil lower member
	QUATERNARY SYSTEM — Pleistocene Series — WISCONSINAN STAGE	soil	soil	soil	Midvale Soil	Toyeh Soil
		alluvium (terrace deposits)	eolian sand, local alluvium	alluvium, loess	eolian sand, local alluvium	Turupah Fm.
10,000	VALDERAN SUBSTAGE	Cochrane till Valders till	Bignell loess Coone alluvial terrace	Pinedale Glaciation: late stade	LAKE BONNEVILLE GROUP: Draper Spit Fm.	LAHONTAN VALLEY GROUP: Sehoo (and Indian Lakes) Fms.: Sehoo Fm., upper mbr.
	TWOCREEKAN SUBSTAGE	Two Creeks forest bed soil	soil	soil	Promontory Point Fm.: soil	Harmon School Soil
	WOODFORDIAN SUBSTAGE	Richland loess Tills from Port Huron thru Shelbyville (30 named units)	Peoria loess (upper part) Ambrose alluvial terrace Tahoka Fm. (upper part), and unnamed basin fills	middle stade	upper member	Sehoo Fm., dendritic mbr. weak soil
20,000		Morton loess		early stade	white marl member	Sehoo Fm., lower mbr.
	FARMDALIAN SUBSTAGE	Farmdale silt and peat Plum Point deposits				
		strong soil	unnamed soil	post-Bull Lake soil	Little Valley Soil	Churchill Soil
30,000	ALTONIAN SUBSTAGE	Winnebago till Roxana silt II to IV	Peoria loess (lower part) Alluvial terrace deposits Tahoka Fm. (lower part) Lake Loman deposits Eolian sands	alluvium, loess	Middle Lake Bonneville alluvium and loess	Wyemaha Fm.
40,000		Winnebago and "Danville" tills Port Talbot deposits		Bull Lake Glaciation: late stade	Alpine Fm. (upper part)	Eetza Fm. (upper part)
50,000		soil		soil?	soil?	soil?
60,000		Dunwich till Roxana silt Ib soil unnamed till(?)		middle stade	Alpine Fm. (middle part)	Eetza Fm. (middle part)
					soil?	soil?
70,000		Roxana silt Ia Scarborough beds		early stade	Alpine Fm. (lower part)	Eetza Fm. (lower part)
break	SANGAMONIAN STAGE	Sangamon Soil alluvium, loess	Sangamon Soil alluvium, loess	Pre-Bull Lake soil Cedar Ridge - Bull Lake Interglaciation	Dimple Dell Soil Younger pre-Lake Bonneville alluvium, colluvium, & loess	Cocoon Soil Paiute Fm.
Time scale arbitrary, and much compressed from the Wisconsinan above	ILLINOIAN STAGE	Buffalo Hart till Jacksonville till & outwash Roby silt Mendon till Petersburg Silt	Loveland Silt Crete sand (gravel) soils?	Cedar Ridge Glaciation	Younger pre-Lake Bonneville lacustrine unit	Younger pre-Lake Lahontan lacustrine unit
	YARMOUTHIAN STAGE	Yarmouth Soil alluvium, loess	Yarmouth Soil alluvium, loess Pearlette ash	soil Sacagawea - Cedar Ridge Interglaciation	pre-Dimple Dell soil Older pre-Lake Bonneville alluvium, colluvium, loess Pearlette ash	pre-Cocoon Soil Older pre-Lake Lahontan alluvium and colluvium
	KANSAN STAGE	Till and outwash Pro-glacial silts Sankoty Sand	Meade Fm. Kansas till Tule Fm.; Hardman alluv. terrace Atchison Fm.	Sacagawea Glaciation	Older pre-Lake Bonneville lacustrine unit	Older pre-Lake Lahontan lacustrine unit
	AFTONIAN STAGE	Afton Soil alluvium, loess	Afton Soil alluvium, loess	soil Dinwoody Lake - Sacagawea Interglaciation		
	NEBRASKAN STAGE	Till and outwash	Blanco Fm. Nebraska till David City Fm.	Dinwoody Lake Glaciation		

CHAPTER VII

APPLICATIONS OF SOIL STRATIGRAPHY TO SOIL SCIENCE

Differences between geologists and soil scientists in their concepts of soils and landscape surfaces

Because of their differing backgrounds and objectives, geologists and soil scientists have some interesting differences in their concepts of soils and landscape surfaces. It may be worthwhile to point out these differences, not because one group is right and the other wrong, but in order to facilitate reaching a fuller, mutual understanding between the two groups.

Differences in pedologist's and geomorphologist's concepts of erosional and depositional surfaces

Pedologists differ from geologists and geomorphologists to rather surprising degree in their concept of erosional and depositional surfaces. Geologists and geomorphologists generally define and delineate such surfaces in terms of form, agency of formation, and relative elevation on a comparatively broad basis. A single valley slope may include surfaces and underlying surficial deposits of widely varying age; this also commonly is true of other geomorphic features such as stream terraces, floodplains, and pediments. The older erosion surfaces are even more broadly defined: ancient "peneplains" such as the Harrisburg Peneplain or the Rocky Mountain Peneplain include a huge assortment of smaller geomorphic features that range very widely in relative elevation and in age. Pedologists, however, adopt a much more restricted viewpoint of surfaces; for them, addition or removal of as little as a foot of surficial material may mean the difference between one surface and another.

Differences between the geologist's concept of weathering profiles (geosols) and the soil-scientist's concept of pedologic units such as soil series

A geosol differs importantly in concept and use from pedologic soil units such as soil series. Geosols are defined primarily on the basis of their stratigraphic and age relations to associated deposits (rock-stratigraphic units) and geomorphic surfaces, not primarily on the basis of profile morphology. Geosols typically are widely traceable, through many variants in profile morphology (soil facies) that would be considered to be separate soil series (or great soil groups) by soil scientists. Only one geosol of a given age is defined for a single area, and this unit is considered to be the same geosol wherever it occurs, no matter how many soil facies it has or how severely it is eroded, provided that enough of its profile remains for it to be identified.

In selecting a type locality for a geosol, a geologist seeks to find a site that displays it as well preserved, with as much of its original profile morphology as possible, in its most typical facies developed under normal environmental (parent material, slope, etc.) conditions; this site also should demonstrate the stratigraphic position of the geosol. The geologist tries to find exposures that are as little modified by later erosion and surficial processes, including soil development, as possible. For the older soils suitable type-locality sites commonly are difficult to find. Buried occurrences generally are better than relict ones, because they are less apt to have been subjected to secondary modification. Rarely, if ever, are the geologist's type-locality sites the same ones that a soil scientist would select for types of soil series units. Soil series types are selected on a quite different basis, to represent average (typical) development of the whole soil series as it is mapped over its whole areal extent.

Rarely is a soil series type representative of a given geosol. In many cases it does not display, reasonably well preserved, the original profile morphology of a geosol--the original profile has been eroded or secondarily modified. Many soil series include more than one geosol, with a younger geosol superposed on an older one. Such relationships are properly designated by soil scientists as unconformities within the soil profile (indicated by Roman numeral horizon designations), but by no means are all such relationships adequately recognized by soil scientists. Soil scientists generally describe a composite or compound(ed) soil profile as a single soil series; geologists differentate the component geosols if they are separately indentifiable.

Another difference arises because to a geologist a soil is a weathering profile; to a soil scientist it is an earth material capable of supporting plant growth, with its lower limit determined by the depth of normal root growth. In some cases geosols may extend well below the depth of root penetration, as in the very deep weathering profiles typical of humid tropical and subtropical regions. In other cases the profile of a weakly developed geosol may not reach the full depth of root penetration. Some completely unweathered deposits, such as very young alluvium, are commonly considered to be soils (with just C-horizon profiles) by soil scientists, yet they obviously are not geosols.

Geologic evidence on time and climate as weathering factors

The factors of weathering profile formation

The various factors that result in formation of weathering profiles are essentially the same as those recognized by soil scientists as the factors of soil-profile formation. The soil-forming factors are generally stated as: climate, time, biotic forces (floral and faunal), topography (slope or relief, and drainage), and parent material (which includes not only the mineral composition, but also geologic factors such as structural relations, bedding, schistosity, etc.). However, in regard to weathering as contrasted with all kinds of soil-profile formation, there are vast differences in rates of action and in relative importance of the different profile-forming factors. Furthermore, the divergences between general soil-forming and weathering processes are much wider in temperate regions than in the tropics. This is mainly because the climatic factor, particularly temperature, is considerably intensified in the tropics. Many of the weathering factors are interdependent, for instance, biotic forces are largely controlled by climate. Because of varying degrees of interdependency, the values of specific factors and their interrelations with other factors commonly are obscure. Quaternary geologic studies are especially suited to give information not only on parent material but also on the two weathering factors that are generally considered to be the most important: time and climate.

Geologic evidence on the time factor in development of weathering profiles

Soil scientists commonly assume, either tacitly or explicitly, that the time factor for a given soil is the "age of the land" (e.g., Joffe , 1949, p. 125), meaning the time that has elapsed since the youngest deposit exposed at the land surface was laid down, or since the youngest erosion surface at the site was formed. A few soil scientists, notably Thorp, Johnson, and Reed (1951), Ruhe (1956), Ruhe and Cady (1958), Ruhe and Daniels (1958), and Butler (1959) and his associates in Australia, have correctly interpreted the significance of buried soils as true time indicators--they obviously formed prior to burial and the difference between the age of the youngest deposit on which they formed and the age of the covering deposit gives the maximum possible duration of the interval when the soil developed its diagnostic characteristics. Quaternary geologic studies have confirmed and supplemented this interpretation, proving that the buried soils are not just accidental occurrences resulting from intermittent local burial during a generally continuous soil-forming climate.

Stratigraphic investigations have been made in a wide variety of climatic and lithogenetic surficial terrains in sufficient detail to bracket narrowly the stratigraphic position of the various weathering profiles, by the methods outlined in chapter 4. They have proved that in every case these profiles developed their diagnostic characteristics during certain specific and widely separated intervals of time regardless of whether they now occur buried or relict.

In addition to the relative chronology established by the physical stratigraphic record, an approximate absolute chronology has been developed, that is based mainly upon radiocarbon dating. As discussed in the section on correlation, the Quaternary geosols are para-isochronous and can be correlated across the U. S. on a stratigraphic basis. Thus determination of the duration of the interval when a given geosol formed in any part of the temperate U. S. should give a close approximation of its duration everywhere in this climatic zone. Table 6.2 summarizes available chronologic information within the range of the radiocarbon dating method and shows current conclusions on the duration of the various weathering intervals (weathering optima) in the Lake Lahontan, Lake Bonneville, Rocky Mountain, Great Plains, and eastern Midwest areas. The weak Early Recent geosol formed probably in less than 200 years, during the interval that Antevs (1948, 1952, 1955) has termed the "Long Drought". The geosol that commonly is called informally the "altithermal soil" (the Toyeh-Midvale-post Pinedale soil) is well dated by radiocarbon and archeologic evidence as having formed within a span of probably not more than 1,000 years in the later part of the altithermal (thermal maximum) interval. The Harmon School-Graniteville-Brady geosol formed during the warmest part of the Twocreekan substage, in considerably less than 1,000 years. The Churchill-Little Valley-post-Bull Lake-Farmdale geosol formed in an interval between 2,000 and 4,000 years long, during the Farmdalian substage. The Sangamonian Stage, when the Cocoon-Dimple Dell-pre-Bull Lake-Sangamon geosol formed, is beyond the range of radiocarbon dating, but the duration of its weathering interval is estimated by other means (such as extrapolated deposition rates and potassium-argon chronology) as between 10,000 and 50,000 years. The Yarmouth soil-forming interval is thought to have been somewhat longer. On the basis of this relatively limited evidence, there appears to be a tendency for the duration of the weathering optima to be somewhat proportional to the degree of development of a given geosol.

The successions in each of these varied lithogenetic and climatic terrains (and also many others) show that the time intervals between these main weathering optima were almost devoid of weathering phenomona. It is true that in some areas weak weathering profiles, preserved as buried geosols, formed during very brief intervals that can be demonstrated to be very small parts of the intervals between main weathering optima. The stratigraphic record clearly demonstrates, however, that nearly all the time between the main weathering optima was characterized by instability, commonly with marked erosion and(or) deposition, and that chemical weathering was too limited to develop even weak weathering profiles.

It is significant that, in comparing the time-spans of the weathering and non-weathering intervals during the last 40,000 years (within the range of radiocarbon dating), the various weathering optima lasted approximately from 1/7 to 1/12 (average about 1/10) as long as the immediately preceding instability interval. This is a much smaller proportion of late Quaternary time than is generally assigned to the weathering optima (for example, see fig. 3.4).

In conclusion, on the basis of stratigraphic records and radiocarbon chronology, it is obvious that the Quaternary weathering optima, when the geosols formed, occupied only brief parts (probably totaling only about 1/10) of all Quaternary time. Each weathering optimum was preceded and followed by relatively longer intervals that were essentially devoid of chemical weathering, at temperate latitudes. In other words, the rate of chemical weathering must have varied greatly, probably by several orders of magnitude. My own interpretation of the variation in weathering rates with time in the Lake Lahontan and Lake Bonneville areas is given in figures 5.1 and 5. 3. (These graphs are only rough estimates because of the many factors involved and the difficulty of obtaining quantitative information on them.)

These conclusions as to the brevity and infrequency of the weathering optima, and the markedly accelerated rate of weathering that they reflect, all suggest that a weathering profile reaches equilibrium with its environment rather quickly. It may then "sit" without much further modification until the climate changes, when under some conditions it may undergo further modificiations.

<u>The climatic factor in weathering profile development</u>

In the past, soil scientists have generally assumed that the zonal soils we see today exposed at the land surface were formed under climate like now; in other words, they assumed that climate has been rather constant in the past, and that differing degrees of soil-profile development are due largely to differing duration of total exposure of the land surface to this climate.

This assumption was made tacitly, mainly for lack of a hint that stratigraphic and paleoclimatic evidence argue to the contrary. Studies in Quaternary geology show, however, that climate did not remain constant but changed continually, and that many of the climatic changes resulted in intervals of climate markedly different from present climate--at times much colder and wetter, at other times warmer and drier, warmer and wetter, etc. Thus it is not safe to assume, as Jenny (1941, p. 180) has stated, that the effect of the longer climatic cycles is relatively insignificant in terms of soil-profile development, and that present climate is a nearly true measure of the climate under which a given soil profile has formed.

That climate today is very unlike that under which the principal geosols formed can be seen by noting the negligible weathering profile development during the Recent epoch (as defined for the Great Basin region by Morrison, 1961d), when climate fluctuated within a small range of present values (Antevs, 1948, p. 179-182, 1952, p.104-106). This epoch started about 4,000 to 3,800 years ago, yet the only discernible weathering profile developed during this whole time was the weak geosol that is called the L Drain Soil in the Lake Lahontan area and generally the "early Recent" soil elsewhere. This minor weathering profile formed during the warmest (and generally driest) part of this epoch, the "Long Drought" of Antevs (1955), which lasted less than 200 years. In short, we are forced to conclude, from the trivial amount of weathering during the last four millenia that is manifest in areas as widely separated as the Great Basin and the eastern Midwest, that geosols like the mid-Wisconsin (Churchill to Farmdale) and altithermal geosols could not possibly have formed under today's climate. Yet the weathering optima when these geosols formed lasted less than 4,000 years! Obviously, therefore, climate must have been significantly different--no other weathering factor has the required periodic variability, yet regional uniformity, to account for the accelerated rate of weathering that is indicated for these optima.

The climate factor has two main components: precipitation and temperature. There seems to be a general assumption among soil scientists that precipitation is the more important of the two, perhaps because its influence on vegetation is particularly obvious, and also there seems to be a correlation between great soil groups and present precipitation values, as Jenny (1941) and others have observed. Adequate precipitation also is a prerequisite for the eluviation-illuviation process in developing soil horizons.

The geologic record, however, challenges this assumption. The wettest parts of the Quaternary, particularly insofar as ground moisture is concerned, were the glacial and lacustral maxima, when the glaciers were extensive and the pluvial lakes were at high levels. (Note, for example, the climate graph for the Lake Lahontan area, in fig. 5.1.) But no evidence has been found of appreciable weathering profile development at these times, in any of the areas where surficial deposits have been studied stratigraphically, either within or contiguous to glaciated areas or far from them. In other words, geologic evidence seems to be negative that increased precipitation alone was capable of producing the accelerated chemical weathering demanded to develop the geosols.

A variety of data point to the other main component of the climatic factor, temperature, as the dominant cause of the accelerated weathering during the weathering optima. When the geosols formed deglaciation was complete and pluvial lakes were dry or at very low levels; the fossil flora (e.g., pollen) and fauna reflect warmth, the pedocal-pedalfer facies change occurs at considerably higher altitudes in the mountains than it does for modern soils (see fig. 4.8); and moreover, the pronounced brown, reddish brown, or reddishB horizons of the stronger geosols, compared with Recent soils,indicate more active hydrolysis and oxidation and a lesser degree of hydration of iron hydrolyzates, also suggests a considerably warmer climate than in Recent time. The more modern stratigraphic studies all correlate the weathering optima with interglacial (or main interstadial) intervals and the more detailed studies correlate them with the later parts of these intervals, which were the warmest (note, for example, Richmond's interpretation shown in fig. 3.4).

The geologic record indicates, however, that strong weathering did not take place during very arid but warm intervals (in regions situated geographically to have such intervals during their climatic cycles). The importance of precipitation relative to temperature is well demonstrated in a study of the climatic history of the Carson Desert, in the Lake Lahontan area, Nevada (Morrison, unpublished manuscript[1/]), which is summarized in the climate graph in fig. 5.1. (This area now is arid, with between 4.5 and 6 inches of mean annual precipitation on the lowlands.) Preliminary values for three important weathering optima in this area are as follows:

	M. a. temperature (°F)		M. a. precipitation (inches)	
Weathering optimum	Actual value	Difference from present value (50.5°F)	Actual value	Difference from present value (5.5")
Cocoon geosol (pre-Lake Lahontan	61	+10.5	8.	+2.5
Churchill geosol (mid-Lake Lahontan)	58	+7.5	8.	+2.5
Toyeh geosol (Altithermal)	56.	6.0	7.+	+2.

On the climate graph in fig. 5.1 it is important to note that the fluctuations in temperature and precipitation were not 100 percent out of phase with each other; in other words, the so-called pluvial intervals were not

1/ Summarized in appendix B.

just cold-wet and the interpluvials just warm-dry. Instead, a more complex general pattern of climatic change is evident: from cool-dry to warm-dry, warm-wet, cool-wet, cold-wet, and back to cool-dry. The pluvial intervals actually started at the beginning of the weathering intervals, when precipitation increased while temperature still was high. The higher temperature _plus_ the greater-than-now precipitation apparently triggered the accelerated chemical weathering that resulted in formation of the geosols. The fact that the weathering optima were parts of the pluvial intervals is not generally recognized because the pluvial lakes remained at low levels, on account of the high evaporation under the higher temperature. (It is indeed possible that the precipitation maximum may have occurred during the weathering optimum, instead of at the time of the maxima (highest levels) of the pluvial lakes.) The significance of this observation becomes more apparent when it is recalled that Bryan and Albritten (1943), Moss (1953), and Butler (1958, 1959) all have correlated the main parts of weathering maxima with glacial-pluvial maxima.

Summarizing, from the Carson Desert climatic study it seems likely that the times of strong weathering were the result of the coincidence of relatively high temperature with at least moderately high precipitation because of somewhat out-of-phase relations of temperature and precipitation fluctuations. In other words, the precipitation was rising during the later part of a thermal optimum, and the time of strongest weathering (in the Carson Desert) was the comparatively brief interval when temperature was still high and precipitation had increased sufficiently for chemical weathering to take place rapidly.

The Carson Desert study also sheds light on the minimum amount of precipitation needed in order to have a weathering optimum. A certain minimum amount of precipitation is necessary before chemical weathering can take place; this can be termed the critical precipitation value. The critical precipitation actually is a range, depending on temperature. This is because the evapotranspiration rate increases with temperature, so that more precipitation is needed to maintain a given soil-moisture content at higher temperatures. A large excess of moisture over this minimum does not seem appreciably to increase the rate of weathering profile development, although it may modify its type. From the Carson Desert study the lowest value of the range of critical precipitation appears to be about 7 inches mean annual precipitation, or an increase of about 50 percent over present precipitation on the lowlands of this area.

This study also suggests that whenever temperature increased above a certain minimum value, which can be called the critical temperature, the rate of weathering profile formation increased markedly, provided that precipitation was at or exceeded its critical value. The critical temperature value seems to be about 52°F. Unlike precipitation, however, as temperature increases above its critical value the rate of profile formation increases exponentially. The explanation for this is not simple, but probably involves at least the following considerations:

Empirically it can be stated that, near room temperature, for every 10°C. rise in temperature the velocity of a chemical reaction increases by a factor of from two to three. This "rule" holds both for numerous chemical reactions, particularly slow ones such as those involved in weathering (particularly hydrolysis), as well as for many biological phenomona. Another reason for the exponential effect of temperature increase on weathering profile development may be because both chemical reactions and biotic activity are affected,particularly activity of micro-organisms. A more important consideration is that ground temperature, rather than mean annual air temperature, is involved. A small increase in mean annual air temperature causes a relatively large increase in the proportion of the year that the soil zone is at or above the critical ground temperature. Vagler (1933) estimates that just the higher soil temperature throughout the year in the tropics increases the chemical weathering rate 2 to 4 times above that in the temperate zone, and in the humid tropics, with high rainfall in addition to year-around high temperature, the rate of chemical decomposition may be increased 20 to 30 times.

Because the rate of weathering profile formation increases exponentially with increase in temperature, temperature is believed to be considerably more important than precipitation in controlling the rate of weathering. Given the minimum moisture (critical precipitation) necessary for active profile formation, temperature appears to become the dominant component of the climatic pedogenic factor. In many cases it probably ranks above time as the chief one of all the weathering factors.

The fact that temperature fluctuations are far more regular in geographic scope (probably synchronous throughout the world), means that the Quaternary geosols were both more ubiquitous in their development, as well as synchronous, than they would have been if precipitation were a controlling factor in their formation. This, in essence, explains why geosols can be considered to be nearly time-parallel (para-isochronous) and thus valid as time-stratigraphic markers.

Limited non-time-parallelism of geosols in going from high to low latitudes,and also from low to high altitudes, is indicated by the following considerations. (This is the reason why they are called "para-isochronous".) Temperature fluctuations are greater in amplitude in temperate and high latitudes than in low latitudes. We can therefore expect that the climatic difference of the weathering optima was proportionally most accentuated in temperate and high latitudes and least so in subtropical and tropical latitudes. This partly reduces the effect of the increase in latitude in shortening the duration of the weathering optima, so that the climate of these times was most different from the climate of the intervening intervals in temperate latitudes, thus making the geosols most conspicuous and effective as time-stratigraphic markers in these latitudes. With decrease in latitude the duration of the weathering optima probably increases, but is less contrasted with average climate, owing to the lower amplitude of the temperature fluctuations (in addition to the higher general annual level of ground temperature). This probably makes geosols less "sensitive" as time-stratigraphic markers at lower latitudes.

Furthermore, in arid areas, such as the Carson Desert, weathering optima probably started somewhat later than they did in more humid areas. During the warm-dry intervals that preceded the warm-wet ones when the geosols formed, the aridity probably was too stark in the arid areas for appreciable chemical weathering to occur. Given weathering optima probably also ended sooner in high mountains than in low-lying areas, because only the peaks of the thermal optima produced sufficient temperature increase to overcome the generally cooler climate in the mountains.

How a soil survey can be helped by an understanding of the Quaternary stratigraphy (including soil stratigraphy) and geomorphology of an area

Every soil surveyor knows that topography and slope conditions commonly have a direct and important influence on soil series. Not so generally recognized, however, are the benefits to be gained from a working knowledge of the geomorphology and Quaternary stratigraphy of an area.

As Woolridge (1949) points out, this kind of geologic knowledge provides much more than a mere schedule of parent materials; its main aim is to discover the evolution of the landscape--the age and genesis of the surfaces of the landscape and of the surficial deposits that underlie them. It is on these surfaces that soils form. The surfaces that need to be considered include not only the older erosional and depositional ones that are marked by benches and platforms, but also the scarp faces, valley sides, and valley floors that have formed recently or are currently in active development. The fact that the surfaces and surficial deposits of a landscape differ in age has been unduly neglected in soil mapping and soil classification.

Geomorphic surfaces and landforms, and the surficial deposits associated with them, are as real and mappable as bedrock units. Commonly they are much more obvious and readily mappable than the latter for a person with adequate geomorphic training. A soil survey of a given area commonly can be expedited if it is preceded by a competent geomorphic study. The boundaries of the main geomorphic surfaces and surficial depositional units generally will be found to constitute the dominant lineaments on the soil map; these boundaries commonly will coincide with those of soil series or of groups of related soil series--and they are much more quickly and easily (and commonly more accurately) determined by geomorphic analysis than by usual augering and soil-pit methods. Recognition of the main geomorphically-determined groupings of soil series will help substantially in generalizing and clarifying the almost intolerable burden of local soil-series nomenclature and differentiation.

If the above recommendation is followed, the pedologic and soil-classification aspects of a soil survey will be aided by deeper insight into the genetic factors, particularly the patterns of distribution of various parent materials, the age and duration of the various weathering intervals, the climatic conditions that prevailed during these weathering intervals, and the subsequent climatic and depositional-erosional history that may be responsible for secondary changes in the soil profiles, such as soil unconformities and secondary pedogenic changes.

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APPENDIX A

Soil-profile sections and chemical and mechanical analyses for the main geosols of the Lake Lahontan area, from their type localities in the southern Carson Desert, near Fallon, Nevada.

Note.--Numeral-and-capital letter notations at the beginning of each soil-profile description [e. g., 32 (S)] identify similar descriptions and also the sample sites given in Morrison, in press, a. Analytic data are from Springer (1953), with minor revisions (such as calculation of calcium carbonate equivalent) by Morrison.

Toyeh Soil

Soil-profile section 32 (S), at type profile of the Toyeh soil, sampled and described by R. B. Morrison and M. E. Springer. Location: About 300 ft E. of road on E. side of Rainbow Mountain, in NE1/2 sec. 17, T. 18 N., R. 31 E. Topographic position: Top of lake bar (highest shoreline of the middle Sehoo lake). Slope: 1 percent. Erosion: Very slight. Exposure: Dug pit. Altitude: 4,176± ft. Parent material: Sand of the dendritic member of the Sehoo formation. Overlying material: None. Vegetation: Greasewood and shadscale, about 6 ft apart; some *Artemesia spinescens*.

Depth (inches)	Thickness (inches)	Soil Horizon	Description
-0.8 to 0	0.75	(lag gravel)	Fine gravel and loose sand. Abrupt, smooth boundary.
0-3	3	A	Pinkish gray (7.5 YR 7/2) very fine gravelly sandy loam; structure, vesicular, weak medium platy; consistence, hard, brittle, harsh; friable; nonsticky, nonplastic. Abrupt, smooth boundary.
3-7	4	B	Light reddish brown (5 YR 6/3) gravelly sandy loam; structure, weak medium to coarse columnar to medium subangular blocky; consistence, hard, slightly harsh; friable; nonsticky, nonplastic. Some soft $CaCO_3$ segregation in lower part. Clear, smooth boundary.
7-14	7	Cca	Light red-brown (5 YR 6/3) gravelly sand; structure, coarse weak columnar; consistence, soft to slightly hard; friable; nonsticky, nonplastic. Slight $CaCO_3$ accumulation (effervesces slightly to moderately with dilute (HCl).
14-35	21	C	Pinkish-gray (5 YR 6/2) gravelly sand; structure, single-grain; consistence, loose; nonsticky, nonplastic.

Table 8.1. **Toyeh soil, soil-profile 32(S), analyses of chemical and physical properties** (analyses by M. E. Springer, Division of Soils, Univ. of California, Berkeley, Calif.)

Soil horizon	Depth inches	Apparent density	Per- cent < 2mm	pH	Percent N (whole soil)	C/N	Calcium carbonate equivalent (percent) (whole soil)	Particle size distribution. Weight-percent < 2mm (organic matter and carbonate-free soil): 2.0-1.0 mm	1.0-0.5 mm	0.5-0.25 mm	0.25-0.10 mm	0.10-0.05 mm	0.05-0.002 mm	< 0.002 mm
A	0-3	1.34	97.1	8.8	0.013	7.0	8.	2.5	6.1	12.1	39.9	13.3	14.4	11.7
B	3-7	1.47	96.3	8.8	0.016	7.3	5.	1.8	3.9	9.2	41.2	12.9	6.4	24.6
Ca	7-14	1.44	95.5	8.9	0.012	6.9	5.7	2.1	3.6	8.6	52.2	13.9	6.2	13.4
C	14-24	1.16	92.9	9.0	0.009	8.6	7.5	2.2	3.3	8.3	59.2	13.7	6.9	6.4
C	24-35	1.43	93.9	8.9	0.009	8.4	5.							

Churchill Soil

Soil-profile section 34 (S), at type locality of Churchill soil (sampled and described by M. E. Springer and R. B. Morrison). Location: East bank of wash gully on west side of Churchill Valley, SE1/4SW1/4 sec. 15, T. 18 N., R. 30E. Topographic position: Steep bank of small mountain wash. Exposure: Vertical channel dug into bank. Altitude: 4190 ft. Parent material: Eolian sand of the Wyemaha formation. Overlying material: Churchill soil is buried under 15 to 20 ft of sand and gravel of the Indian Lakes and Sehoo formations. Note. The soil profile here is slightly truncated, with removal of at least 4 inches of the uppermost part of the original profile.

Depth (inches)	Thickness (inches)	Soil Horizon	Description
-4-6	10	B2	Light brown (7.5YR6/3) medium sand with sparse rock fragments; structure, moderate very coarse angular blocky; consistence, hard, firm; nonsticky, nonplastic. Clear, smooth boundary.
6-12	6	B3	Light brown (7.5YR6/3) medium sand with a few white lime concretions; structure, almost massive (very weak, very coarse angular blocky); consistence, hard; friable; nonsticky, nonplastic. Clear, smooth boundary.
12-92	80	Cca	Very pale brown (10YR7/3) medium sand with white lime streaks, "concretions", and irregular concentrations; structure, structureless, massive to single-grain; consistence, very hard and very firm in upper part, ranging to loose in lower part. $CaCO_3$ concentration decreases somewhat irregularly from top to bottom; upper 2 ft have numerous to common lime streaks, concretions, etc., and are massive and locally almost cemented, and very hard to hard; remainder of thickness has some to few lime concentrations, is single-grain and slightly hard to loose, except for a 1/2 in. to 1 in. white $CaCO_3$ cemented layer at 49-50 in. depth. Diffuse boundary.
92-112	20	C	Light gray medium sand (10YR7/2) with sparse white lime streaks and concretions; structure, single-grain; Consistence, loose.

Table 8.2. Churchill soil, soil-profile 34(S), chemical and physical properties

(analyses by M. E. Springer, University of California, Berkeley, Calif.)

Depth Inches	Apparent Density	Per-cent <2mm	pH	Percent N (Whole Soil)	C/N	Calcium carbonate equivalent (percent) (Whole Soil)	Particle size distribution Weight-perct <2mm (organic matter and carbonate-free soil)						
							2.0-1.0 mm	1.0-0.5 mm	0.5-0.25 mm	0.25-0.10 mm	0.10 0.05 mm	0.05 0.002 mm	<0.002 mm
							1.3	44.1	33.7	17.9	1.7	0.8	0.5
0 - 6	1.69	98.6	8.6	0.012	9.6	0.7	0.9	8.5	16.5	44.0	12.3	10.0	7.8
6 - 12	1.50	97.8	8.6	0.009	8.7	3.	0.5	10.9	19.4	37.8	10.9	4.6	15.9
12 - 18	1.45	97.8	8.9	0.005	--	9.	0.6	20.8	21.3	35.2	9.3	4.1	8.7
18 - 24	1.47	96.9	8.9	0.005	8.8	8.							
24 - 36	1.53	98.3	9.1	0.002			0.4	16.5	26.2	42.0	8.2	3.0	3.7
36 - 48	1.56	99.4	9.1	0.002		0.7							
50 - 62	1.56	99.9	8.9	0.002		0.2							
62 - 80	1.49	99.8	8.8	0.002		0.2							
80 - 92	1.55	100.0	9.0	0.002		0.4							
92 - 112	1.53	100.0	9.2	0.001		1.							

L-Drain Soil

Soil-profile section 36 (S), at the type locality of the L-Drain soil; described and sampled by R. B. Morrison and M. E. Springer. Location: Northwestern part of Nevair Flat, NE1/4 sec. 16, T. 18 N., R. 29 E. Topographic position: Nearly level plain. Exposure: Dug pit. Erosion: none. Altitude: 3,939 ± 3 ft. Parent material: Alluvial sand coeval with the first lake unit of the Fallon formation. Overlying material: 1 1/2 in. of eolian sand of the upper mbr. of Fallon formation.

Depth (inches)	Thickness (inches)	Soil Horizon	Description
-1.5-0	1.5		Sand, single-grain, loose (overlying material).
0-1	1	A	Brown (7.5YR5/2) sandy loam; structure, weak vesicular, very weak medium platy (almost massive); consistence, soft to slightly hard, very friable.
1-3	2	A	Brown (7.5YR5/2) sandy loam; structure, weak medium granular; consistence, friable, slightly hard.
			Abrupt, smooth boundary.
3-7	4	B	Brown (7.5YR5/3) loamy sand; structure, very weak medium granular; consistence, slightly hard.
7-13	6	Cca(?)	Brown (7.5YR5/2) sand; structure, very weak medium granular; consistence, nearly loose; effervesces slightly with dilute HCl.

Table 8.3. L-Drain soil, soil-profile 36(S), chemical and physical properties
(Analyzed by Dr. M. E. Springer, Division of Soils, University of California, Berkeley.)

Soil horizon	Depth inches	Apparent density	Percent < 2 mm	pH	Percent N (whole soil)	C/N	Calcium carbonate equivalent (percent) (whole soil)	Particle size distribution Weight-perct < 2 mm (soil free of organic matter and carbonate)						
								2.0-1.0 mm	1.0-0.5 mm	0.5-0.25 mm	0.25-0.10 mm	0.10-0.05 mm	0.05-0.002 mm	0.002 mm
A (vesicular)	0-1	1.51	91.1	9.6	0.006		1.							
	1-3	1.46	94.9	9.5	0.006	10.0	0.7	11.2	8.2	7.8	34.1	20.1	8.3	10.3
B	3-7	1.51	99.6	9.5	0.006	8.9	0.7	13.1	18.2	10.2	26.1	16.3	6.9	9.2
Cca(?)	7-13	1.58	95.7	9.6	0.006	6.9	0.5	17.6	28.4	12.0	22.7	9.2	3.6	6.5
C	13		87.5	9.6	0.005			26.1	31.4	13.4	16.3	5.9	3.4	3.5

Cocoon Soil

Soil-profile section 58 (S), of the Cocoon soil, sampled and described by M. E. Springer and R. B. Morrison. Location: Summit plateau of White Throne Mountains, 15 miles south of Fallon, Nev., NE1/4NW1/4 sec. 15, T. 16 N., R. 29 E. (1/4" SW of T of Mountains, in Carson Lake quad.) Topographic position: Broad, nearly flat ridge crest. Exposure: Dug pit. Altitude: 4,740 feet. Slope: 1 percent. Erosion: Very slight to none. Parent material: Solifluction and creep mantle about 15" thick, underlain by vesicular olivine basalt of the Bunejug formation. Natural cover: Sparse shrubs (Shadscale, 30 percent ; little greasewood, 60 percent), very sparse *Bromus tectorum*.

Depth (inches)	Thickness (inches)	Soil Horizon	Description
-1 to 0	1±	--[1]	Desert pavement of dark brown varnished flaggy basalt blocks and pebbles, covering 90 percent of surface.
0-2	2	A[1]	Pinkish gray (5 YR 7/2) fine sandy loam; structure, vesicular, moderate coarse columnar; weak medium platy; consistence, slightly hard, harsh, friable.
			Abrupt, smooth boundary.
2-6	4	B2	Reddish brown (5 YR 5/3) gravelly clay loam; structure, moderate medium granular; to weak medium subangular blocky; consistence, slightly hard to hard; slightly sticky, slightly plastic, friable. Contains more roots than layers above and below.
			Clear, irregular boundary.
6-9	3	B3ca	Light reddish brown (5 YR 6/3) gravelly sandy loam; structure, weak, medium granular; consistence, loose. A few roots; numerous $CaCO_3$ concretions below 7 inches.
			Gradual boundary.
9-15	6	Cca	Pink (5 YR 7/3) gravelly sandy loam; contains many flat $CaCO_3$ concretions; structure, massive; consistence, hard to very hard, very firm to extremely firm; weakly to strongly cemented.
			Gradual boundary.
15		Cca	White $CaCO_3$ cementing fractured dark gray basalt.

[1] Probably has formed later than the Cocoon soil proper and hence not properly a part of its profile.

Table 8.4. Cocoon soil, soil-profile 58(S), chemical and physical properties
(analyzed by M. E. Springer, Univ. of California, Berkeley, Calif.)

Soil horizon	Depth inches	Apparent density	Percent < 2 mm	pH	Percent N (whole soil)	C/N	Calcium carbonate equivalent (percent) (whole soil)	Particle size distribution Weight-perct < 2 mm (O.M. and carb. free soil)						
								2.0-1.0 mm	1.0-0.5 mm	0.5-0.25 mm	0.25-0.10 mm	0.10-0.05 mm	0.05 0.002 mm	< .002 mm
A	0 - 2	1.33	97.2	9.3	0.017	8.3	3.2	0.5	0.3	0.6	18.6	26.7	39.2	14.1
B2	2 - 6	1.04	99.8	7.6	0.047	12.5	0.2	0.3	0.4	0.6	14.1	19.8	21.5	43.3
B3ca	6 - 9	1.35	57.2	8.4	0.026	10.4	31.	1.1	3.3	4.5	26.0	24.5	28.1	12.5
Cca	9 -15	1.47	21.5	8.5	0.014		53.	2.6	2.0	3.6	25.7	23.4	22.7	20.0
	15		0.8											

Soil-profile section 59 (S), of the Cocoon soil, showing especially well preserved upper part of profile (sampled and described by M. E. Springer and R. B. Morrison). Location: Top of White Throne Mountains, 15 miles south of Fallon, Nevada. On line between sec. 15 and 16, T. 16 N., R. 29 E., at southern margin of Carson Lake quadrangle. Topographic position: Crest of ridge. Exposure: Dug pit. Altitude: 4,860 ft. Slope: ± 1 percent. Erosion: Very slight to none. Parent material: Colluvium (solifluction and creep mantle) about 1 1/2 feet thick, underlain by olivine basalt of the Bunejug formation. Present climate: Average mean annual precipitation about 6 inches; average mean annual temperature about 49°F. Natural cover: Sparse, covers less than 3/4 of surface; mainly shadscale (*Atriplex confertifolia*), little greasewood (*Sarcobatus Baileyi*), but sage (*Artemesia spinescens*), and rarely other shrub species, together with sparse grasses (mostly *Bromus tectorum*) and forbs. Most of the bare part of the surface is covered by a desert pavement of slabby blocks and pebbles of basalt, only one stone thick; generally oriented parallel with the surface. The top surfaces of the stones are mostly shiny dark brown, dark reddish brown, to nearly black due to desert varnish, whereas the under surfaces are dull brownish gray or gray.

Depth (inches)	Thickness (inches)	Soil Horizon	Description
Top			
-1-0	1±	--	Desert pavement of dark brown varnished, flaggy blocks and pebbles of basalt, covering 90 percent of surface.
0-2.5	2.5	A[1]	Pinkish gray (7.5 YR 7/2) very fine sandy loam; structure, vesicular (numerous spherical or tubular voids 1/4 to 3 mm in diameter) moderate coarse columnar, weak medium platy; consistence, slightly hard, harsh, floury; friable. Cracked vertically to form polygonal blocks 2.5 to 4 in. in diameter.
			Abrupt, smooth boundary.
2.5-7	4.5	B2	Reddish brown (5 YR 5/4) gravelly clay loam; structure, moderate to strong medium granular to moderate subangular blocky (upper 1/4 inch is strong fine-granular); consistence, slightly hard, friable; slightly sticky, slightly plastic.
			Clear, irregular boundary.
7-10	3	B3ca	Light reddish-brown (5 YR 6/4) gravelly clay loam; structure, granular to nuciform; consistence, slightly hard, friable, slightly sticky, slightly plastic. Some $CaCO_3$ concretions.
			Clear, wavy boundary.
10-14	4	Cca	Pink (5 YR 7/3) gravelly sandy loam, strong $CaCO_3$ cementation; structure, weak granular; consistence, very hard.
			Gradual boundary.
14-19	5	Cca	Pink, (5 YR 8/3) stony sandy loam, some $CaCO_3$ cementation; structure, weak granu- consistence, very hard, extremely firm; weakly to strongly cemented.
			Gradual boundary.
19-60	41	R	Fractured basalt, cemented with $CaCO_3$ (soil lime).

[1] Probably younger than the Cocoon soil and hence not properly a part of its profile.

Table 8.5. Cocoon soil, soil profile 59(S), chemical and physical properties
(analyzed by M. E. Springer, Univ. of California, Berkeley, Calif.)

Soil horizon	Depth inches	Apparent density	Percent < 2 mm	pH	Percent N (whole soil)	C/N	Calcium carbonate equivalent (percent) (whole soil)	Particle size distribution Weight-percent ≤2 mm. (O.M. and carb. free soil)						
								2.0-1.0 mm	1.0-0.5 mm	0.5-0.25 mm	0.25-0.10 mm	0.10-0.05 mm	0.05-0.002 mm	< .002 mm
A	0 - 2.5	1.52	99.1	9.0	0.012	9.7	1.6	0.5	0.5	0.7	22.8	26.7	35.9	12.9
B2	2.5 - 7	1.03	93.3	6.4	0.026	12.6	0.2	0.5	0.6	1.0	14.9	20.0	22.7	40.3
B3ca	7 - 10	1.12	50.3	7.3	0.014	11.0	1.8	5.5	4.8	4.7	26.2	21.8	22.9	14.1
Cca	10 - 14	1.56	47.8	8.7	0.011	10.0	10.	4.4	9.0	8.7	36.0	23.0	15.6	3.3
Cca	14 - 19		15.4	8.6	0.004		22.							
	19 - 36		22.3	8.5	0.005		12.	7.2	8.3	10.0	30.9	20.3	15.9	7.4
	36 - 60						8.5							

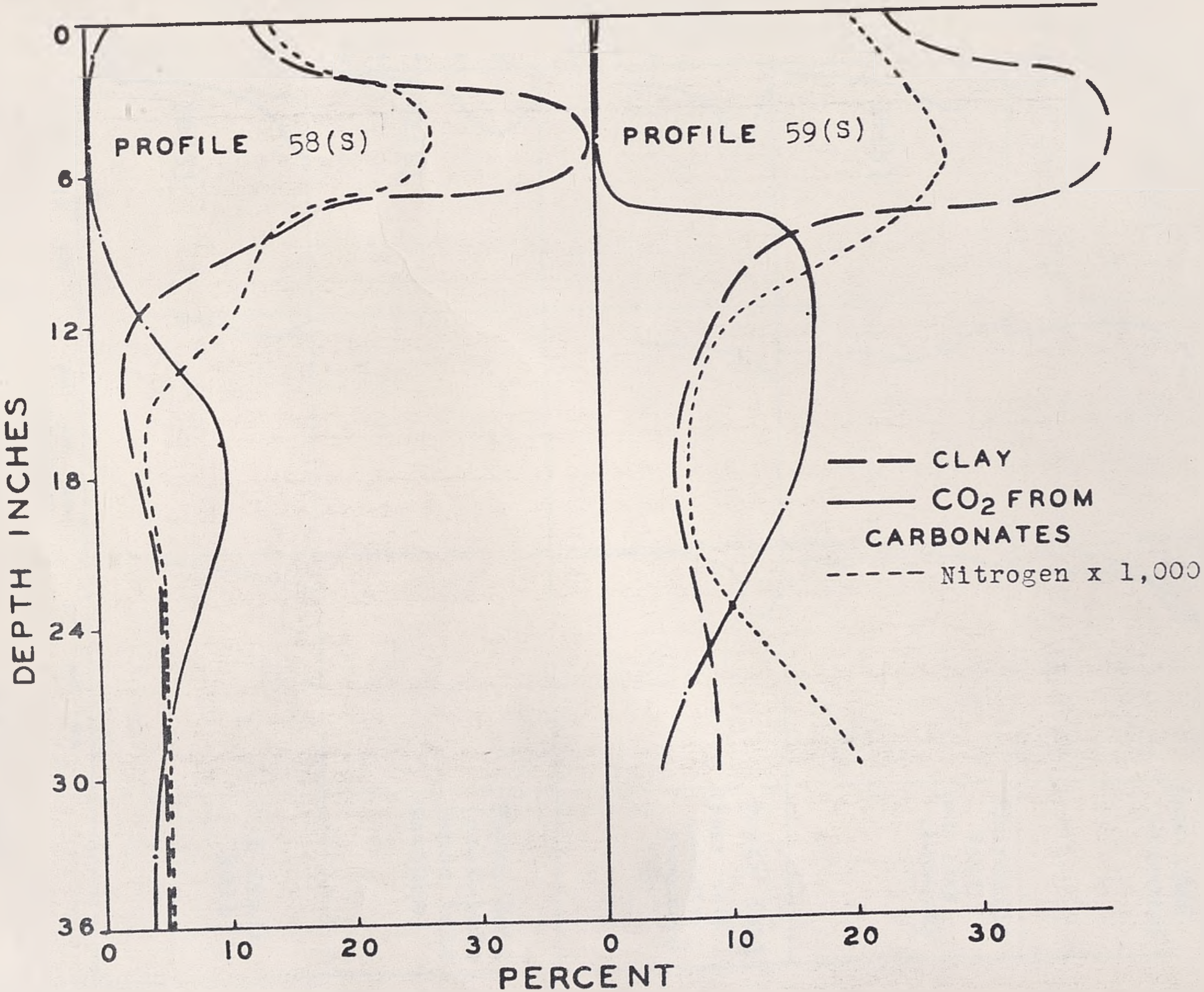

Fig. 8.1. Distribution of soil properties in two profiles of the Cocoon Soil.

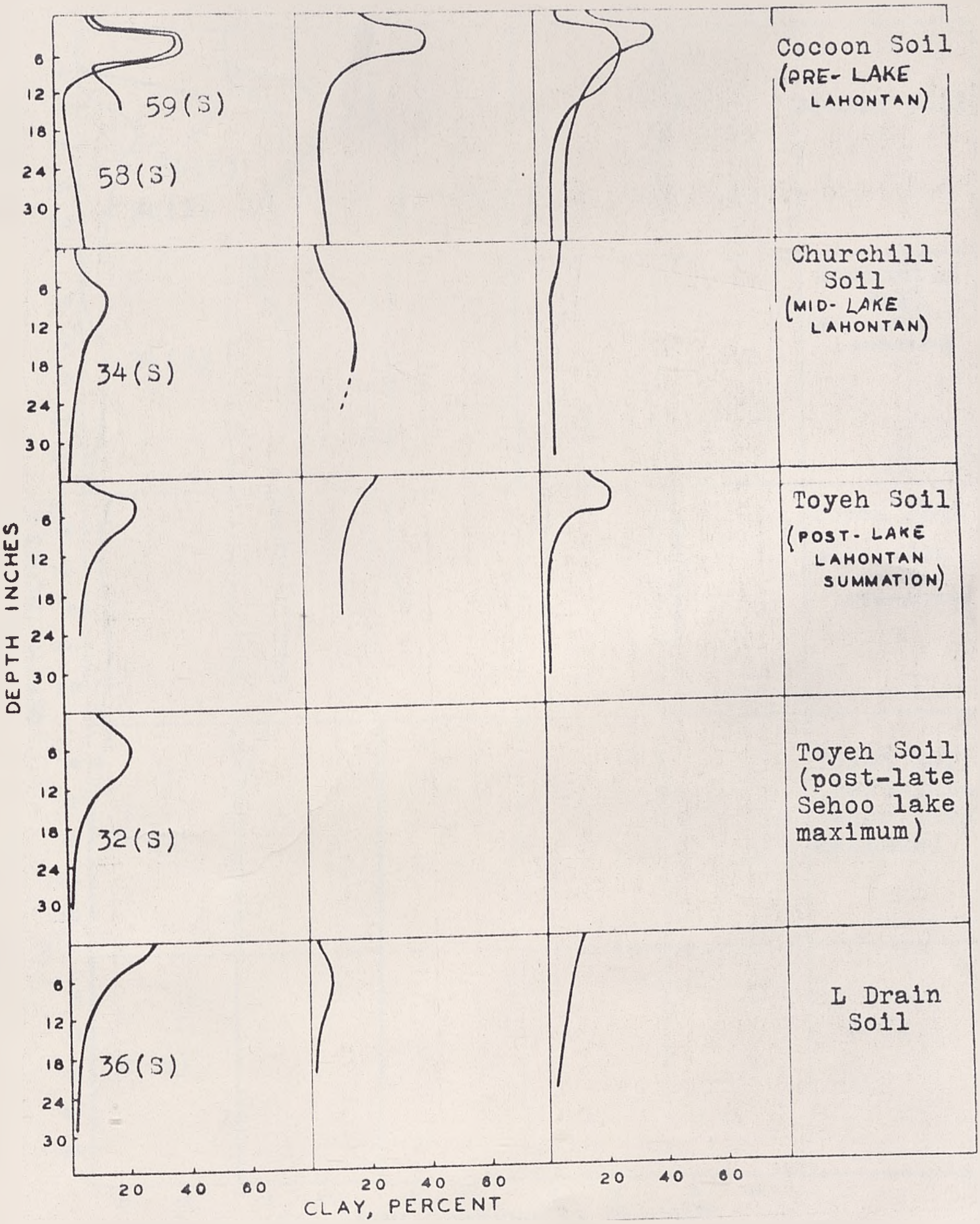

Fig. 8.2. Distribution of the 0.002 mm. (clay) fraction in the profile. Based on <2 mm. soil freed of organic matter and carbonates.

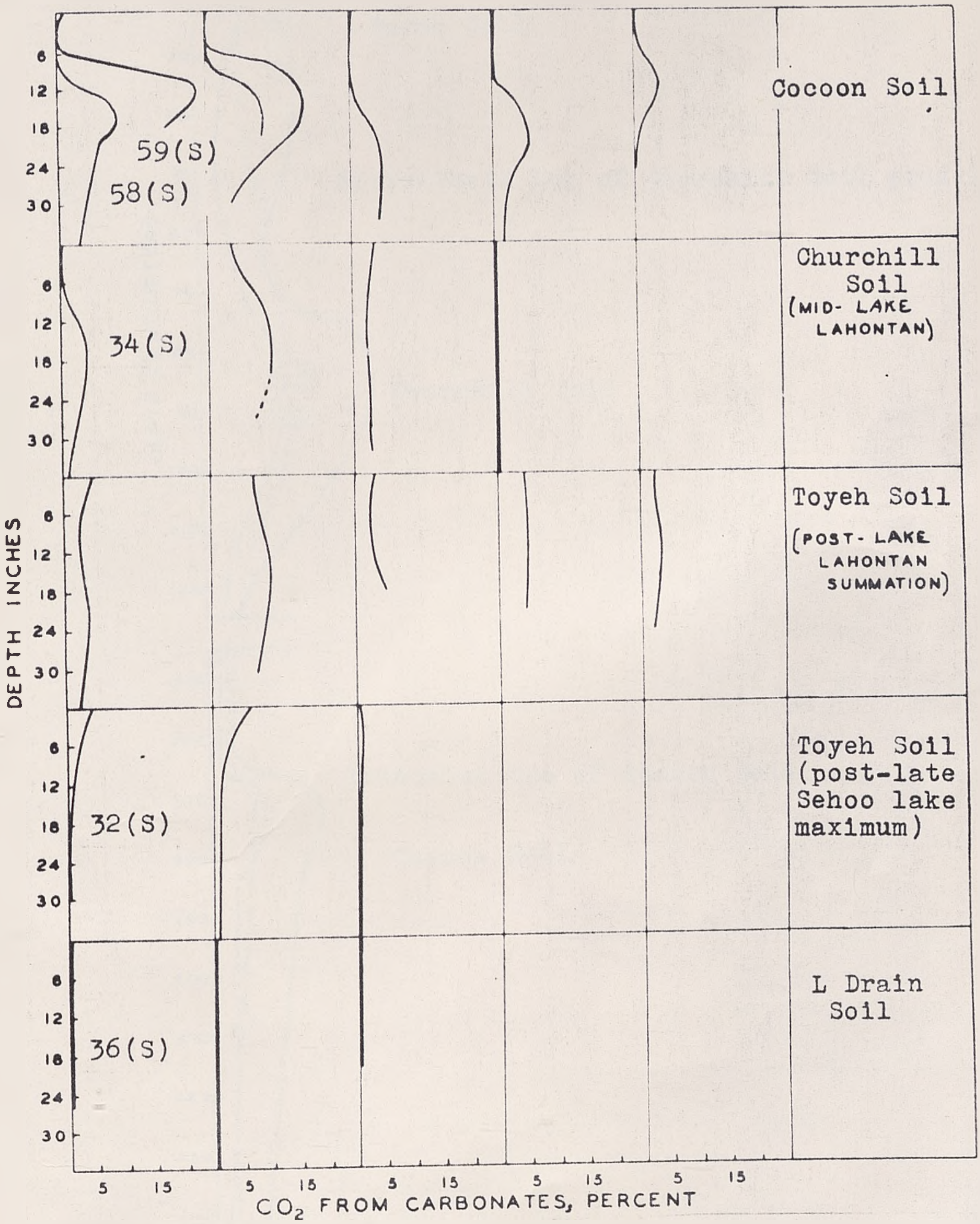

Fig. 8.3. Distribution of CO_2 from carbonates in the profiles. Percentages are based on the whole soil (includes 2 mm. as well as 2 mm. fractions).

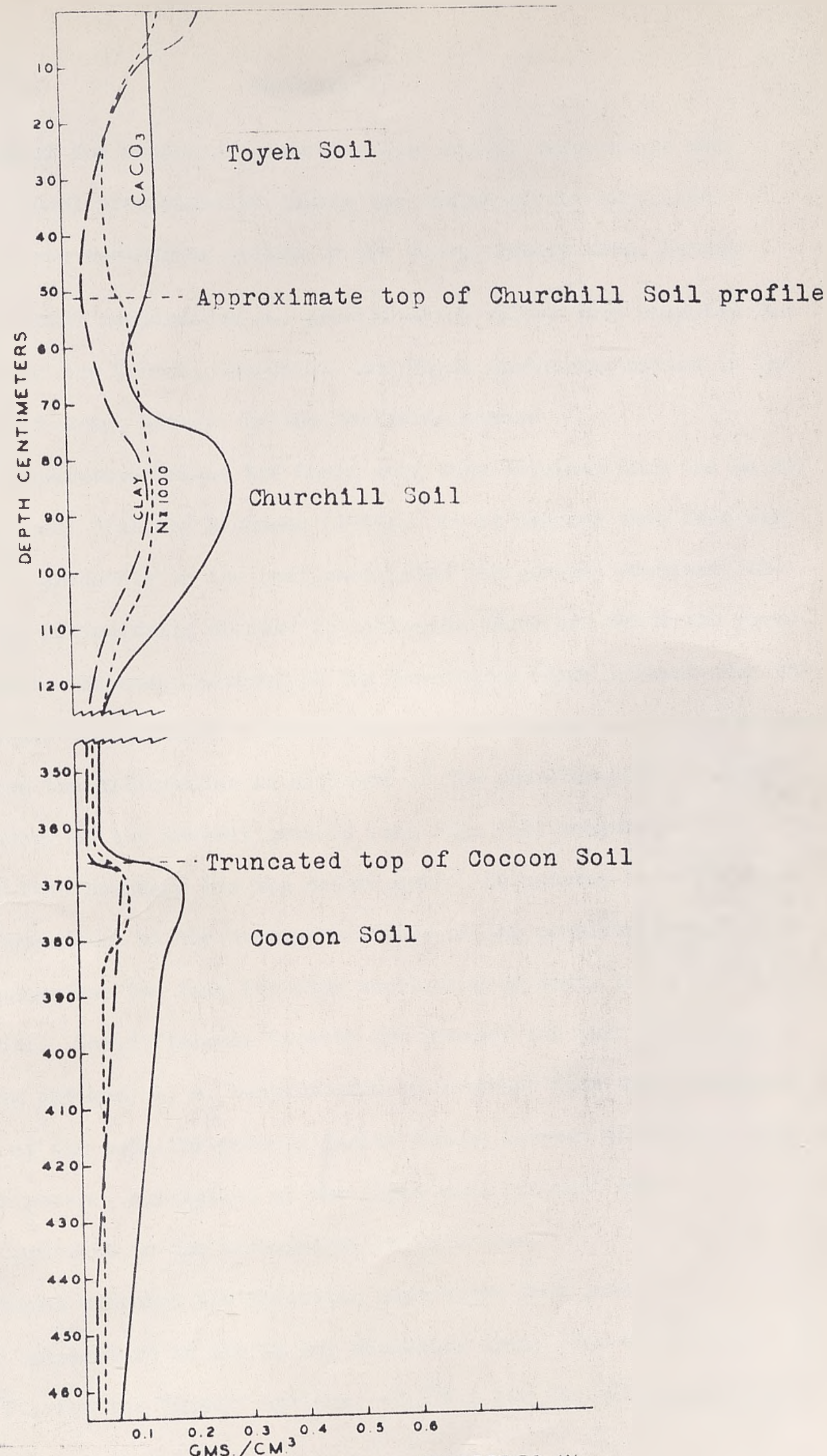

Fig. 8.4 DISTRIBUTION OF SOIL PROPERTIES IN A THREE-STORY PROFILE

APPENDIX B

Resume of the method of computing mean annual temperature and mean annual precipitation during the maxima of the main late Quaternary weathering optima in the Carson Desert area, Nevada

Mean annual temperature and precipitation values were computed for the maxima of the Cocoon, Churchill, and Toyeh weathering optima in the Carson Desert area, Nevada, by the following means:

(1) Temperature values are based upon data obtained from the La Sal Mountains area, Utah, by Richmond (1953). It is assumed that free air temperature gradients in the past paralleled the present gradient, although they periodically shifted to altitudes above and below the present gradient. Richmond determines the departures (from present mean annual free-air temperature) of mean annual temperatures during the weather-maxima from the differences in altitude of the pedalfer-pedocal soil facies boundary of the ancient geosols (see fig. 3.6) compared with the altitude of this boundary for the modern soil. He assumes that the mean annual temperature at the average altitude of the pedalfer-pedocal soil facies change was the same for each weathering optimum, including the modern soil. The difference between the present and past (for a given weathering optimum) m. a. temperatures is obtained from the difference in altitude of the pedalfer-pedocal facies change between the modern soil and the given geosol, multiplied by the lapse rate (present mean vertical temperature gradient in the atmosphere) in this area.

Richmond computed the following departures from present mean annual free-air temperature in the La Sal Mountains area: for the Sangamon optimum, 10°F.; for the "Brady" optimum, +7.5°F.; for the altithermal optimum, +5.8°F.

These differences between past and present mean annual (m. a.) free-air temperatures can be transferred to the Lake Lahontan area without correction because the lapse rate is essentially the same in the two areas, as evinced by data on the lapse rates at Reno, Nevada, and at Grand Junction, Colorado. Therefore, Richmond's computed temperature departures can be transferred directly from the La Sal Mountains to the Lake Lahontan area, by merely adding them to the curve of the present m. a. free-air temperature gradient (with altitude) in the latter area, to establish another curve showing the m. a. free-air temperature gradient in the Lake Lahontan area during a given weathering maximum. From this past temperature gradient curve the past m. a. temperature (during the weathering maximum) at any altitude in the Lake Lahontan area can be read. On this basis, the m. a. free-air temperature departures (from present values) for maxima of weathering optima in the lowlands of the Carson Desert, Nevada, are as follows: for the Cocoon maximum (correlative with the Sangamon maximum), +10.5°F.; for the Churchill maximum (correlative with Richmond's "Brady" maximum), + 7.5°F.; for the Toyeh maximum (correlative with the altithermal), + 6.0°F.

(2) Mean annual precipitation values for the maxima of the weathering optima are determined by using the above temperature values indirectly for determining the equilibrium evaporation-runoff relations for near-desication lake levels in the Carson Desert. (The geologic record shows that very low-level lake conditions existed during the weathering optima in this basin, which was the sump for nearly half of the total drainage of Lake Lahontan.) It is possible to calculate either the mean annual temperature or the m. a. precipitation that will provide the runoff needed to maintain a terminal lake of a given area, if the other factor is known.

Past values of m. a. precipitation are first computed as average values over the whole Lake Lahontan drainage basin. These values cannot be calculated directly, owing to several variable factors, but are arrived at by a series of approximations. The steps in doing this are as follows:

(A) Assume a given value (in inches) for the average m. a. precipitation over the whole drainage area of the Carson Desert (P_b), for a given weathering maximum.

(B) Compute the precipitation on the lake area (assumed near dessiccation) that existed at this time by solving the equation:

(Equation 1) $P_l = \dfrac{P_b \times P'_l}{P'_b}$, where

P_l = past m. a. precipitation on the lake area, in inches; and

P'_l = the average between the present m. a. precipitation at the altitude of the given lake level, in inches, and the present m. a. precipitation at the floor of the Carson Desert, in inches;

P_b = past average m. a. precipitation, in inches, for the whole drainage basin; and

P'_b = present average m. a. precipitation, in inches, for the whole drainage basin.

P_l always is less than P_b because precipitation decreases with altitude, and the lake level in the terminal basin always is below the effective mean altitude of the drainage area.

(C) The runoff needed to maintain the lake is expressed by the following equation:

Equation 2) $R = \dfrac{A_l\,(E_l - P_l)}{A_n}$, where

R = m. a. runoff from the drainage area (A_n), in inches;

A_l = area of the lake, in square miles (determined from a lake-area curve, for various depths of the lake);

A_n = area of the drainage basin, in square miles;

E_l = m. a. evaporation (gross) from the lake, in inches; and

P_l = m. a. precipitation on the lake, in inches.

The lake evaporation can be determined from the graph in Langbein and others, (1949, fig. 2), showing the relation between mean annual temperature and mean annual free-water evaporation, using the temperature value (determined as discussed above) for the given weathering maximum at the effective mean altitude of the drainage area.

(D) Read P_b from the graph in Langein and others (ibid.), using the above-determined R value and the known or assumed m. a. temperature at the effective mean altitude of the drainage area.

(E) Repeat as many times as needed, until P_b as calculated in step (D) equals the value assumed for P_b in step (A).

To convert the Pb values (average m. a. precipitation for the whole Carson Desert drainage area to precipitation values at other altitudes, the following equation is used:

(Equation 3) $$P_a = \frac{P_b \times P'_a}{P'_b}$$, where

P_a = past m. a. precipitation at the given altitude, in inches;

P'_a = present average m. a. precipitation at the altitude zone of which the given altitude is the mean, in inches;

P_b = past average m. a. precipitation for the whole drainage basin, in inches; and

P'_b = present average m. a. precipitation for the whole drainage basin, in inches.

For converting the P_b values for the whole Carson Desert drainage basin to precipitation in the lowlands of the Carson Desert (average altitude 4,100 feet),

P'_a = 5.5 inches and P'_b = 10.5 inches, so that $P_a = P_b \times 0.57$.

By the above-described means, the precipitation values for the Cocoon, Churchill, and Toyeh weathering maxima that are given on page 135 were obtained.