SOILS AS A TOOL FOR APPLIED QUATERNARY GEOLOGY

by

Peter W. Birkeland, Michael N. Machette, and Kathleen M. Haller

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UTAH GEOLOGICAL AND MINERAL SURVEY
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UTAH DEPARTMENT OF NATURAL RESOURCES

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INTRODUCTION

This manual provides a brief summary of materials discussed during a short course entitled "Soils as a tool for applied Quaternary geology" that was taught May 30-June 1, 1990, under the auspices of the Utah Geological and Mineral Survey's Mineral Lease Program. This course consists of a 1½-day seminar and a 1-day field trip to selected localities in the Salt Lake City area.

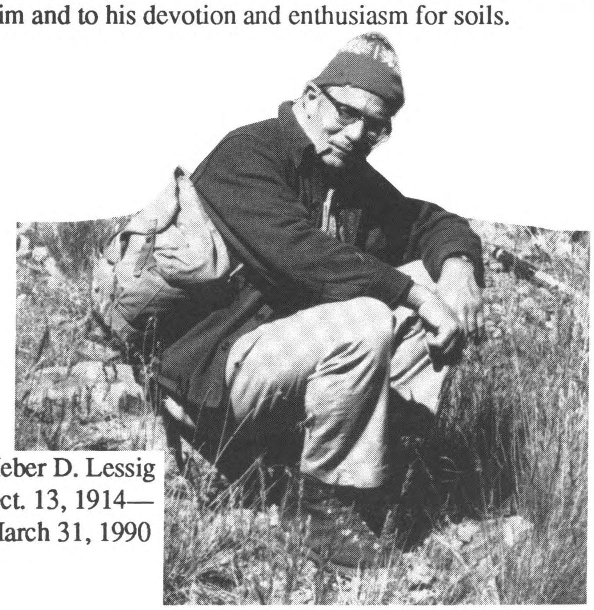
During the seminar, we introduced geologists to the field of pedology as it applies to Quaternary geology and geomorphology. Pedology (the study of soils) and soil stratigraphy (the relation between soils and geologic units) are scientific specialities that are poorly appreciated and rarely applied to studies in Utah. This situation has arisen for three reasons: (1) soil stratigraphy and pedology are specialized fields that bridge agronomy and geology; (2) the academic community, being strongly compartmentalized, generally does not teach graduate-level courses that combine these subjects; and thus, (3) many locally trained scientists have no practical experience with these techniques. Unfortunately, only a few universities in the Western United States have such capabilities.

Soils are ubiquitous features of modern and ancient landscapes. It has long been recognized that the degree of soil development is, in part, a function of time and, as such, soil development can be a relative measure of the age of the parent materials. In addition, soils commonly have morphological, mineralogical, and geochemical signatures of the climate(s) under which they formed and, thus, can be indicators of paleoclimatic conditions.

The quantitative assessment of soils is both a useful relative-age technique and a paleoclimatic indicator that has widespread applicability to a broad number of scientific and applied geologic problems throughout the State of Utah. Soils are easily described in the field (*i.e.*, a mapping tool) by using a standardized system of diagnostic characteristics and nomenclature. The use of quantitative measures of soil development that are based on field characteristics, such as the profile-development index, can provide geologists with estimates of relative ages of 10 to 500 ka for a variety of surficial materials. Further quantification by laboratory analyses, some of which are relatively simple and inexpensive (such as total content of secondary calcium carbonate), can provide a basis for correlation with dated deposits on both a local and regional basis. Soils can provide relative age estimates beyond the limits of radiocarbon dating (commonly 30,000-40,000 yr), which is the most widely used numerical-dating technique in Quaternary geology.

Soils are used in geologic mapping to assign a relative sense of age to stratigraphic frameworks (rather than just young to old) and to indicate the duration of unconformities in stratigraphic sequences. In applied geology, soils can help to assess the age and stability of landforms (piedmont surfaces, fault scarps, landslides, *etc.*). The seminar and this manual are intended to introduce soils as a mapping and quantitative tool for geologists. We present an overview of these topics and review several case studies from Utah and the Western United States in which soils were used to solve problems in applied geology.

Finally, a note about Heber D. Lessig who passed away recently. Heber was a former Soil Conservation Service Soil Scientist, Utah resident, and friend of Quaternary geology. Birkeland first got in contact with Heber in about 1973. Heber had experience mapping soils in Ohio and he contributed data on a soil chronosequence along Little Beaver Creek, Ohio (Lessig, 1961), for Birkeland's (1984) book on soils. Upon retirement, Heber and his wife, Bonnie, moved to Salt Lake City where Heber continued to work on soils when they were not skiing or travelling. Heber attended many of the Friends of the Pleistocene trips, and was nicknamed the "human pipette" for his amazingly accurate estimates of soil texture. His knowledge of soils and his good humor always made us look forward to our next trip with him. Once, he and Bonnie put up the entire Colorado FOP contingent in their house! Heber's latest contribution was a descriptive report on soils in the Wasatch Front of Utah under contract to the Utah Geological and Mineral Survey's Mineral Lease Program. We had planned to use Heber's soil descriptions and have him lead the field part of this short course. Instead, we dedicate this manual and short course to him and to his devotion and enthusiasm for soils.



Heber D. Lessig
Oct. 13, 1914—
March 31, 1990

ACKNOWLEDGMENTS

We thank Jennifer Harden and Dan Muhs for reading an early draft of this manuscript. Some of the research reported on by Birkeland was supported by both a Faculty Fellowship and a grant from the Council on Research and Creative Work of the University of Colorado, and a Gladys W. Cole Memorial Research Award of the Geological Society of America.

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CHAPTER I
THE SOIL PROFILE AND SOIL-HORIZON NOMENCLATURE

A soil is a natural body consisting of layers or horizons of mineral and (or) organic constituents of variable thicknesses, which differ from the parent material in their morphological, physical, chemical, and mineralogical properties and their biological characteristics; at least some of these properties are pedogenic (Fig. 1-1). Soil horizons generally are unconsolidated, but some contain sufficient amounts of silica, carbonates, or iron oxides to be cemented. The study of soils in the field is called pedology, and pedogenic processes are responsible for the form (morphology) of most soils.

THE SOIL PROFILE

A soil profile consists of the vertical arrangement of all the soil horizons from the surface down to the parent material. In studying a soil, therefore, the investigator must be able to identify the parent material from which the soil formed. This is no easy task and requires a good deal of experience in Quaternary geology and pedology. However, once the parent material is recognized and its original properties estimated, one can begin to determine departures from the properties of the original material and identify these materials as soil horizons.

Some geologists distinguish between a soil profile and a weathering profile. Where this is done, the soil profile is generally considered to make up the upper part of the much thicker weathering profile. As one illustration, the soil horizons have lost most properties of rock or sediment (*i.e.*, stratification), whereas the weathering profile may be a saprolite in which the original rock structures are preserved to some extent. For our purposes, we will consider the soil profile as extending to the unaltered parent material.

SOIL-HORIZON NOMENCLATURE

Most Quaternary geologists use the soil-horizon nomenclature of the Soil Survey Staff (1975) with some appropriate modifications. This nomenclature has evolved over the years, the most recent revision having come in 1981. A

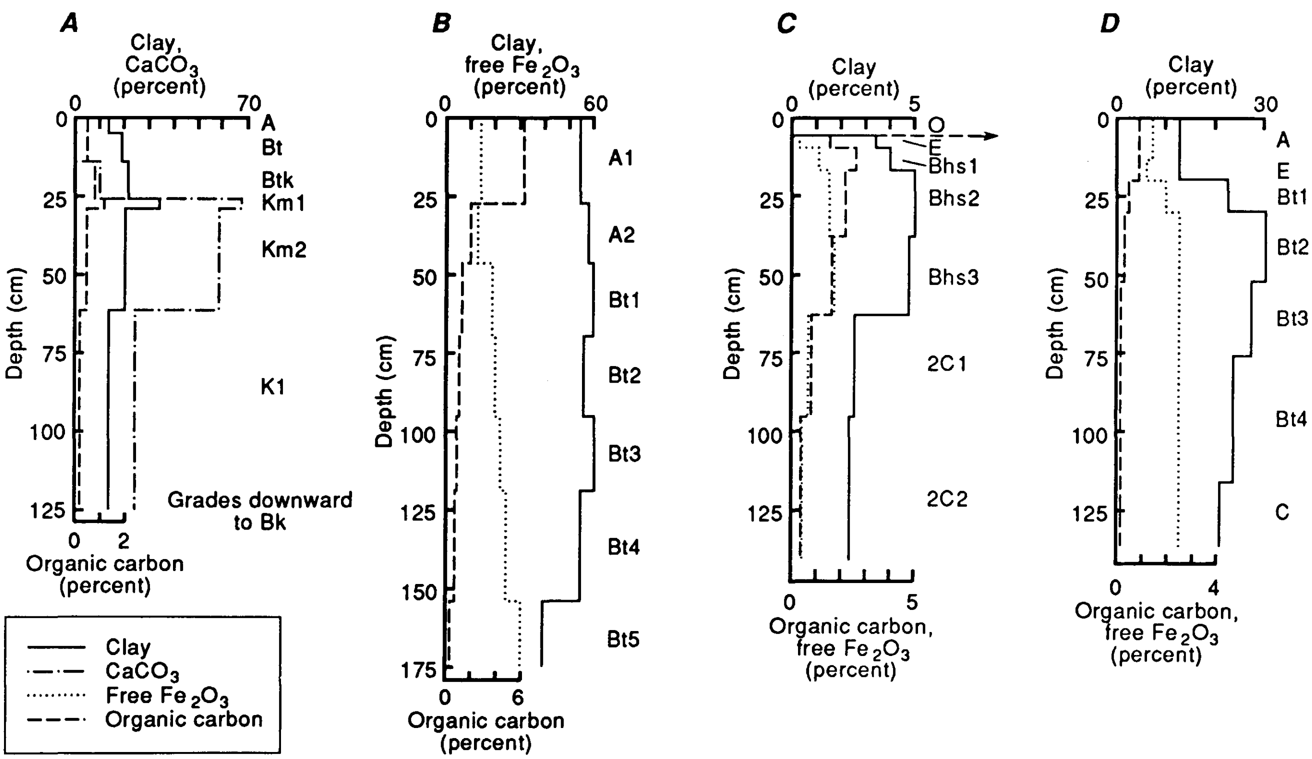


Figure 1-1. Laboratory data on soil profiles that illustrate properties commonly associated with various soil horizons. Percent clay is for the noncarbonate fraction, and both percent clay and carbonate are for the nongravel fraction. ***A***. Parent material is gravelly alluvium (soil at Terino from Gile and others, 1966, Table 2). ***B***. Parent material is serpentine (from Soil Survey Staff, 1960, profile no. 27). ***C***. Parent material is glacial outwash (from Soil Survey Staff, 1960, profile no. 20). ***D***. Parent material is loess (from Soil Survey Staff, 1960, profile no. 11).

comparison between the old (pre-1981) and new field description systems is given in Table 1-1. Soil-profile depth functions are helpful in visualizing the various key properties of the more common horizons (Fig. 1-1). The underlying principle of soil-horizon nomenclature is eluviation-illuviation in the profile. One example is the movement of iron (Fe) in a soil under acid-leaching conditions: Fe is removed from the upper part of the profile (eluvial part) and accumulates at depth (illuvial part). Another example is the movement of carbonate in a semiarid environment: carbonate is leached (eluviated) from the upper part of the profile and is accumulated (illuviated) in the lower part.

Three kinds of symbols are used in soil-horizon nomenclature: capital letters, lower-case letters, and Arabic numerals. Capital letters denote master horizons, lower-case letters denote some specific characteristic or subdivision of the master horizon, and Arabic numerals denote either a further subdivision of the horizon or discrete layering of the parent material.

Most soil profiles can be divided into several master or prominent horizons. Surface or near-surface horizons relatively high in organic matter are designated O or A horizons, the difference between the two being determined by the amount of organic matter present. For the O horizon, the mineral fraction usually is a smaller percentage of the horizon's volume than in an A horizon and is less than about 80% by weight. Beneath the O or A horizon in some environments (commonly in leaching environments typical of coniferous forests), there is a light-colored (E) horizon that is relatively leached of iron compounds. However, in most environments, the B horizon is beneath the surface horizon (or horizons) and in most cases is a horizon of accumulation (illuviation). This horizon encompasses a multitude of soil characteristics relative to those of the assumed parent material. Among the B-horizon characteristics are clay accumulation, the production of reddish colors, the accumulation of iron compounds with or without organic matter, the residual concentration of resistant materials following the removal of more soluble constituents under conditions of intensive weathering and leaching over long intervals of time, and accumulation of $CaCO_3$ and more soluble salts. A slightly weathered C horizon (Cox) commonly is beneath the B horizon and beneath that is either unweathered bedrock of the R horizon, or unweathered, unconsolidated material of the Cu horizon. In desert environments, carbonate buildup plays an important role in soil morphology and genesis, and horizons dominated by carbonate are designated K. If a pedogenic carbonate horizon does not meet the criteria for a K horizon, it is designated Bk.

Table 1-1.—A comparison between the old and new soil-horizon nomenclature
[From Guthrie and Witty (1982)]

MASTER HORIZONS		SUBORDINATE DEPARTURES		
Old	New	Old	New	Description
O	O	—	a	Highly decomposed organic matter.
O1	Oi, Oe	b	b	Buried soil horizon.
O2	Oa, Oe	cn	c	Concretions or nodules.
A	A	—	e	Intermediately decomposed organic matter.
A1	A	f	f	Frozen soil.
A2	E	g	g	Strong gleying.
A3	AB or EB	h	h	Illuvial accumulation of organic matter.
AB	—	—	i	Slightly decomposed organic matter.
A&B	E/B	ca	k	Accumulation of carbonates.
AC	AC	m	m	Strong cementation.
B	B	sa	n	Accumulation of sodium.
B1	BA or BE	—	o	Residual accumulation of sesquioxides.
B&A	B/E	p	p	Plowing or other disturbance.
B2	B or Bw	si	q	Accumulation of silica.
B3	BC or CB	r	r	Weathered or soft bedrock.
C	C	ir	s	Illuvial accumulation of sesquioxides.
R	R	t	t	Accumulation of clay.
		—	v	Plinthite.
		—	w	Color or structural B.
		x	x	Fragipan character.
		cs	y	Accumulation of gypsum.
		sa	z	Accumulation of salts.

In arid regions, the progressive development of carbonate horizons has been classified by division into several morphological stages. Gile and others (1965, 1966) recognized four stages (I-IV) (Fig. 1-2); later, Machette (1985a) further subdivided stage IV for a total of six stages (see Appendix). The stages are identified on particular morphological features that correlate well with duration of soil formation. Gypsum or halite can accumulate under both extreme cold and dry conditions (*e.g.*, Antarctica) or warm aridic climates, and one can use the same six-stage morphological scale as is used for carbonate.

Within the C horizon, numbers have been used to denote variation in properties with depth, but these numbers carry no specific meaning (C1, C2 . . . C8, C9). In the Western United States, many workers find the differentiation into Cox for oxidation and Cu for unweathered horizons useful.

In the field, many horizons are transitional rather than sharp. In places, the transitional material is not described, but is included in the description of the distinctness of the horizon boundary (see Appendix). However, there may be soils in which the transitional material should be described and sampled. There are two kinds of transitional horizons. In one, the properties of both horizons are mixed but one property is dominant; in this case, both capital letters are used and the first letter indicates the dominant properties of that horizon (*e.g.*, AB, BA, AC). The other kind of transitional horizon has distinct parts of both horizons; here the two capital letters are separated by a virgule (*e.g.*, A/B, E/B, B/R).

Master horizons are further subdivided by use of both lower-case letters and Arabic numerals (Table 1-2). The lower-case letters follow the capital (master horizon) letter (*e.g.*, Bt). If it is possible to determine the sequence of pedogenic events, the order of lower-case letters should reflect that sequence; for example, Btk would indicate clay accumulation followed by carbonate accumulation. If the horizon is buried, the b is usually written last (Btkb); however, if a soil property is imparted after burial, the symbol denoting the property follows b (Btbk). Arabic numerals are used to further subdivide horizons identified by a unique set of letters. Such subdivision can be based on slight changes in color, structure, or any other property. The numbering starts at 1 and runs consecutively as long as the capital-letter plus lower-case-letter sequence remains constant. Horizons in a particular profile could be A, AB, Bt1, Bt2, Btk, Bk1, Bk2, Bk3, and Btb, or A, Bt, K1, K2, K3, and K4.

Many unconsolidated deposits of Quaternary age consist of depositional layers of contrasting texture (including variable percentages of gravel and/or different lithologies), and the soil profile extends through more than one layer. Examples are layered sequences of loess/till, flood-plain silt/gravelly outwash, and colluvium/outwash. Such primary differences in texture and (or) lithology are important in any soil-profile description and in determining soil genesis. Each different geological layer is so noted by an Arabic numeral, counting from the top down. In previous classifications, Roman numerals were used for this purpose. The numerals precede the master-horizon designation, and the numeral for the uppermost layer (1) is omitted. A sequence in a soil profile of A, Bt1, 2Bt2, 2Bt3, and 3Bk would indicate three distinct parent materials; the sequential numbering of subhorizons of the Bt indicates that the soil formed simultaneously across the boundary between parent-material units 1 and 2.

Soils can be classified by their position in a stratigraphic section and in the landscape (Ruhe, 1965; Morrison, 1978). For example, three soils are recognized in Figure 1-3. Relict soils are those that have formed on an old surface and have never been buried; they may or may not have acquired most of their properties some time in the distant past. Buried soils are those that were formed on an ancient land surface and were subsequently buried by a younger deposit; many are far enough below the present land surface to not have been significantly affected by contemporary pedogenic processes. If, however, a buried soil is close to the surface, processes affecting the surface soil also will affect the buried soil (Ruhe and Olson, 1980; Schaetzel and Sorenson, 1987). Exhumed soils are soils that were buried, but because of subsequent erosion, they are presently at the surface.

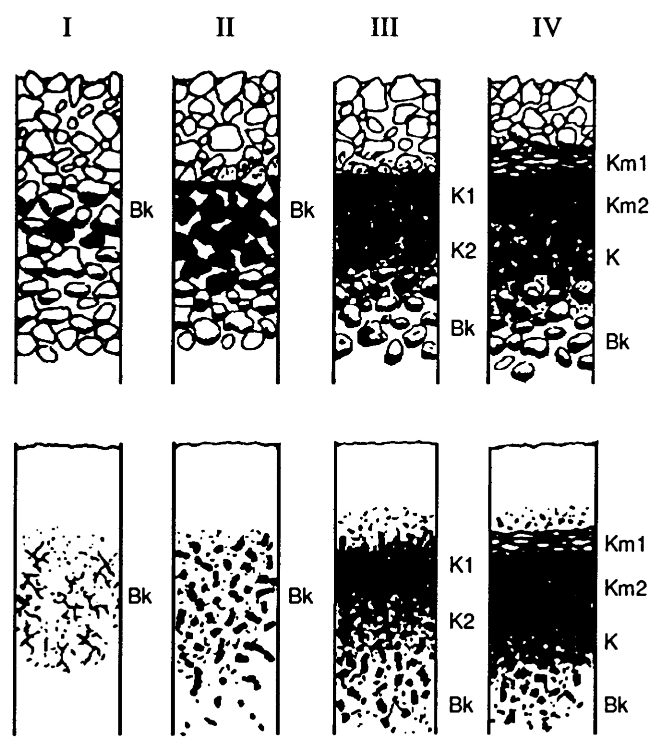


Figure 1-2. Sketch of carbonate buildup stages (I, II, III, IV) for gravelly (top) and nongravelly (bottom) parent materials (from Gile and others, 1966). See Appendix for definitions of these stages and for stages V and VI.

Table 1-2.—Soil-horizon nomenclature
[Modified from Guthrie and Witty (1982) and unpublished (1981) manuscript of revised U.S. Department of Agriculture Soil Survey Manual]

DESCRIPTION OF MASTER HORIZONS AND SUBHORIZONS

O horizon. Surface accumulations of mainly organic material; may or may not be, or has been, saturated with water. Subdivided on the degree of decomposition as measured by the fiber content after the material is rubbed between the fingers:

Oi horizon—Least decomposed organic materials; rubbed fiber content is greater than 40% by volume.

Oe horizon—Intermediate degree of decomposition; rubbed fiber content is between 17 and 40% by volume.

Oa horizon—Most decomposed organic materials; rubbed fiber content is less than 17% by volume.

A horizon. Accumulation of humified organic matter mixed with mineral fraction, and the latter is dominant. Occurs at the surface or below an O horizon; Ap is used for those horizons disturbed by cultivation.

E horizon. Usually underlies an O or A horizon, characterized by less organic matter and/or fewer sesquioxides (compounds of iron and aluminum) and/or less clay than the underlying horizon. Horizon is light colored due mainly to the color of the primary mineral grains because secondary coatings on the grains are absent; relative to the underlying horizon, color value will be higher or chroma will be lower.

B horizon. Underlies an O, A, or E horizon, shows little or no evidence of the original sediment or rock structure. Several kinds of B horizons are recognized, some based on the kinds of materials illuviated into them, others on residual concentrations of materials. Subdivisions are:

Bh horizon—Illuvial accumulation of amorphous organic matter-sesquioxide complexes that either coat grains, form pellets, or form sufficient coatings and pore fillings to cement the horizon.

Bhs horizon—Illuvial accumulation of both organic matter and sesquioxides as organic matter-sesquioxide complexes; both value and chroma are three or less.

Bk horizon—Illuvial accumulation of alkaline earth carbonates, mainly calcium carbonate; the properties do not meet those for the K horizon.

Bl horizon—Illuvial concentrations primarily of silt (Forman and Miller, 1984).

Bo horizon—Residual concentration of sesquioxides, the more soluble materials having been removed.

Bq horizon—Accumulation of secondary silica.

Bs horizon—Illuvial accumulation of amorphous organic matter-sesquioxide complexes; both color value and chroma are greater than three.

Bt horizon—Accumulation of silicate clay that has either formed *in situ* or is illuvial; hence it will have more clay than the assumed parent material and/or the overlying horizon. Illuvial clay can be recognized as grain coatings; bridges between grains; coatings on ped surfaces or in pores; or thin, single or multiple near-horizontal discrete accumulation layers of pedogenic origin (clay bands or lamellae). In places, subsequent pedogenesis can destroy evidence of illuviation.

Bw horizon—Development of color (redder hue or higher chroma relative to C horizon) or structure with little or no apparent illuvial accumulation of material.

By horizon—Accumulation of secondary gypsum.

Bz horizon—Accumulation of salts more soluble than gypsum.

K horizon. A subsurface horizon so impregnated with carbonate that its morphology is determined by the carbonate (Gile and others, 1965). Authigenic carbonate coats or engulfs nearly all primary grains in a continuous medium. The uppermost part of a strongly developed horizon is laminated, brecciated, and (or) pisolithic (Machette, 1985a). The cemented horizon corresponds to some caliches and calcretes.

C horizon. A subsurface horizon, excluding R horizon, like or unlike material from which the soil formed, or is presumed to have formed. Lacks properties of A and B horizons, but includes materials in various stages of weathering.

Cox and Cu horizons—In many unconsolidated deposits, the C horizon consists of oxidized material overlying seemingly unweathered C. The oxidized C does not meet the requirements of the Bw horizon. In stratigraphy, it is important to differentiate between these two kinds of C horizons. We suggest the Cox be used for oxidized C horizons and Cu for unweathered C horizons. Cu is from the nomenclature of England and Wales (Hodson, 1976).

Cr horizon—In soils formed on bedrock, there commonly will be a zone of weathered rock between the soil and the underlying rock. If it can be shown that the weathered rock has formed in place, and has not been transported, it is designated Cr. Such material is the saprolite of geologists; *in situ* formation is demonstrated by preservation of original rock features, such as grain-to-grain texture, layering, or dikes. If such material has been moved, however, the original structural features of the rock are lost, and the transported material may be the C horizon for the overlying soil.

R horizon. Consolidated bedrock underlying soil. It is not unusual for this and the Cr horizon to have illuvial clay in cracks; the latter would be designated Crt.

Table 1-2.—Soil-horizon nomenclature—Continued

SELECTED SUBORDINATE DEPARTURES

Lower-case letters follow the master horizon designation. Those that are mainly specific to a particular master horizon are given above. Some can be found in a variety of horizons; they are listed below.

- b Buried soil horizon. May be deeply buried and not affected by subsequent pedogenesis; if shallow, they can be part of a younger soil profile.
- c Concretions or nodules cemented by iron, aluminum, manganese, or titanium.
- f Horizon cemented by permanent ice. Seasonally frozen horizons are not included, nor is dry permafrost material; that is, material that lacks ice but is colder than 0°C.
- g Horizon in which gleying is a dominant process, that is, either iron has been removed during soil formation or saturation with stagnant water has preserved a reduced state. Common to these soils are neutral colors, with or without mottling. Strong gleying is indicated by chromas of one or less, and hues bluer than 10*Y*. Bg is used for horizon with pedogenic features in addition to gleying; however, if gleying is the only pedogenic feature, the horizon is designated Cg.
- j Used in combination with other horizon designation (Btj, Ej) to denote incipient development of that particular feature or property (National Soil Survey Committee of Canada, 1974).
- k Accumulation of alkaline earth carbonates, commonly $CaCO_3$.
- m Horizon that is more than 90% cemented. Denote the cementing material (Km, carbonate; qm, silica; kqm, carbonate and silica; *etc.*).
- n Accumulation of exchangeable sodium.
- v Horizon characterized by iron-rich, humus-poor, reddish material that hardens irreversibly when dried. Called plinthite (Soil Survey Staff, 1975). If A horizons in arid environments have a vesicular structure (round voids), they are designated Av.
- x Subsurface horizon characterized by a bulk density greater than that of the overlying soil, hard to very hard consistence, brittleness, and seemingly cemented when dry (fragipan character).
- y Accumulation of gypsum.
- z Accumulation of salts more soluble than gypsum (for example, NaCl).

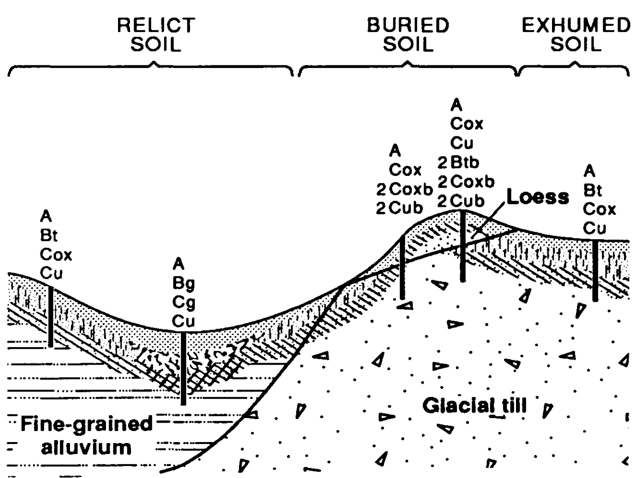


Figure 1-3. Lateral variation in soil profiles due to variation in environmental conditions and lithology and age of parent materials. Soil horizons parallel the land surface. Those horizons that are developed from river alluvium truncate depositional layering; hence, some soil properties are independent of properties of depositional layers. The soil formed on glacial till was partly truncated by erosion before burial by loess; it grades laterally to the right from a buried to an exhumed soil, because the loess cover has been removed by erosion in that direction.

RECOGNITION OF BURIED SOILS

Some buried soils are so obvious that few people would argue as to their pedogenic origin; others, however, are quite difficult to differentiate from geological deposits. At field conferences, many people have had the experience that one person's soil is another's geological deposit. Part of the problem might be that some people have not spent enough time studying surface soils to really understand their properties and profile characteristics.

The same criteria used to recognize and describe surface soils should be used for buried soils (Yaalon, 1971a; Valentine and Dalrymple, 1976). The first test is to trace the material laterally in outcrop to be sure that it is a soil and not a deposit. The relation between the soil horizons and bedding should be deciphered because soil horizons can truncate geological bedding (Fig. 1-3). Also, there are predictable lateral changes in soils related to topographic position (Chapter V), so one will want to be certain that these exist and that they are not lateral changes in parent material. A problem arises where a soil forms on a surface that parallels geological bedding. Common examples are soils formed on alluvial-fan deposits in the semiarid southwestern United States. Whenever deposition stopped for a long enough time, a soil could form, and it could have been buried during subsequent deposition. In this case, the soil horizons would parallel the bedding in the deposit. Some parts of alluvial-fan deposits can be poorly sorted, however, and resemble soils in thickness, color, and texture. Thus, one has to look for features in that zone to be sure soil formation has taken place. Most soil features used to recognize modern soils should persist after burial (Yaalon, 1971b). In general, the organic matter in the A horizon does not persist long after burial (Holliday, 1988), but the mineral part of the A horizon may still be present and recognizable by a slightly lower clay content than that of the B horizon. Generally, the buried B horizon is the most important horizon for recognizing buried soils. If it is a Bt, it will have a greater clay content, redder or browner colors, and have a better developed structure than the C horizon. In drier regions, the presence of a carbonate-enriched horizon beneath the Bw or Bt horizon is also helpful in identifying a buried soil. If a K horizon is present, it should have a morphology similar to that commonly seen in surface soils (Fig. 1-2). Moreover, carbonate accumulation horizons should have a distinct relation to the buried land surface, and to overlying soil horizons (Fig. 1-1*A*), so that a ground-water origin for the carbonate commonly can be ruled out.

One other criterion that is helpful in the recognition of a buried soil is the abruptness of the horizon boundaries (Fig. 1-1). Quite commonly, the upper boundary of a soil horizon is sharper than the lower boundary, but this need not be the case in depositional layering. For example, with depth in a well-developed soil profile the transition from a low-clay-content A horizon to a clay-enriched Bt horizon may take place over a few centimeters, whereas below the clay maxima in the Bt horizon there is a more gradual decrease in clay content toward the C horizon. The same criterion appears to hold for pedogenic carbonate horizons; that is, the upper horizon boundary is sharp and marked by a thin transition zone between the overlying noncarbonate material and the carbonate-enriched horizon, whereas the lower boundary is much less sharp. If a Bw horizon is present in a well-drained soil, the color is redder or browner in the upper part of the horizon, and redness of hue and chroma gradually diminish with depth. In contrast, post-burial diagenetic alteration within a deposit may result in a gradual decrease in color hue or intensity both upward and downward from the zone of maximum alteration and may be localized in materials of particular textures.

Another criterion that is helpful in the recognition of buried soils is the mineralogical characteristics of the profile. Some nonclay silicate minerals become weathered and (or) etched upon weathering and if the zone studied is a B horizon, weathering and etching may be greater there than in the underlying C horizon and the overlying deposit. Mineral depletion during weathering may be reflected by the ratio between resistant and less-resistant minerals that have a consistent relation with depth. Clay minerals may also give a clue on the pedogenic origin of a horizon. Commonly, the clay minerals that form during pedogenesis vary predictably in type with depth in a profile.

A more difficult problem in the recognition of buried soils comes when the upper part of the soil has been removed by erosion, leaving only that part of the profile that was below the original B horizon. Here, oxidation colors help in identifying the material as part of a buried soil, as long as post-burial alteration can be ruled out and the soil was originally well drained. In the midcontinent, evidence for such a history is shown by the carbonate content of superimposed loess sheets. The older loess may have been leached of carbonate during an interval of soil formation, the soil B horizon may have been removed during a subsequent period of erosion, and the leached loess may then have been buried by carbonate-bearing loess (Ruhe, 1968; Willman and Frye, 1970). In this case, the major remaining evidence for an unconformity and an interval of soil formation is the presence of carbonate-bearing loess overlying loess that had been leached of its carbonate prior to burial.

Stone lines may help in the identification and location of buried soils. Stone lines are thin, buried, more or less planar layers of stones (Ruhe, 1959). Although several origins have been suggested for them, including biological activity (Johnson, 1989), some are distinctly of geological origin—such as a lag deposit formed when the associated erosional surface was cut or concentration of stones at the base of a creeping colluvial mantle. Whatever the geological origin, a stone line could be a field indicator of a hiatus in deposition, and if so, it could be associated with a soil or truncated soil.

Soil chemical data can help with the identification and interpretation of buried soils. Constituents that are analyzed, however, should be those that persist after burial (Yaalon, 1971b). Chemical trends in iron, aluminum, and phosphorus probably would persist, but trends in pH and exchangeable cations could be altered soon after burial and give no information about values of the preexisting soil (Ruhe and Olson, 1980). In arid regions, calcium carbonate could be subsequently translocated into a buried Bt horizon to give the common Btbk horizon. In most places where pedogenic carbonate accumulates, the pre-burial carbonate morphology below the pre-burial Bt horizon should be recognized.

The recognition of buried soil horizons could be aided by thin-section studies of soil mineralogy or fabric. In places in the field, it may be difficult to differentiate parent-material features from pedologic ones, and the only real clues that the materials are soils could come from thin-section analysis (Douglas and Thompson, 1985; Kemp, 1985).

CHAPTER II
QUANTIFICATION OF FIELD SOIL DESCRIPTIONS— THE PROFILE-DEVELOPMENT INDEX

INTRODUCTION

The degree of soil-profile development, based on field descriptions, can be used in a qualitative or quantitative manner as a measure of the amount of pedologic change that has taken place in the time since the parent material was deposited. Profile development is a common tool in Quaternary stratigraphy where soils are used to correlate unconsolidated deposits. The qualitative ranking is generally on a relative scale, based on properties of a sequence of soils in an area. Morrison (1978) illustrates how this is done, and Richmond (1962) used the method in his study of the La Sal Mountains of southeastern Utah. However, a quantitative scale is more useful, because soil of the same age may vary in its development from place to place due to variations in other soil-forming factors.

Two quantitative profile-development indices have been suggested recently, one by Bilzi and Ciolkosz (1977) and one by Harden (1982). The latter index is preferred by most workers in the Western United States, probably because all of the soil-horizon properties are compared with the parent material, or assumed parent material, and soil horizon depths are included in the calculation. Because parent-material properties might not be represented at depth in the profile (Cu horizon), one might have to seek them elsewhere. For example, in working with a river-terrace soil chronosequence, one could use the properties of sediment of the present active flood plain as an analog for the soil parent material. The profile-development index of Harden (1982), which we discuss in the following section, is referred to as the PDI.

CALCULATION OF PROFILE-DEVELOPMENT INDEX

There are nine steps in calculating the PDI (Fig. 2-1) as described by Harden (1982) and Harden and Taylor (1983), with some modification.

1. Describe the soil profile. In any study, the same individual should describe all the soils, so that uniform descriptions are obtained. Reheis and others (1989) reported on the variations in PDI that can occur when several individuals describe the same soil.
2. Assess the parent material or parent materials, and describe them in the same way you would a soil horizon (color, texture, clay films, *etc.*). In places, the Cu horizon is the parent material. In other places, weathering extends so deep that unaltered parent material is not reached in the exposure of the soil; therefore, you must find a reasonable substitute (surrogate) for the parent material. For example, modern river alluvium can be used for the parent material of a river terrace, or "Little Ice Age" (latest Holocene) till at the front of glaciers might be used for the parent material for tills farther downvalley. Contemporary flood-plain silt (overbank alluvium) might be representative of the parent material for loess that is flood-plain derived. If, however, the loess is derived from the reworking of an older loess, one may use the properties of the latter. Parent-material properties for colluvium are difficult to ascertain; perhaps one could use the properties of ground-up rock as a parent-material surrogate. Finally, in the unusual case where an older soil is the parent material for a younger soil (*e.g.*, if they formed before and after a significant climatic change, respectively), then the older soil properties are valid for the parent-material properties of the younger soil.
3. Quantify the degree to which the soil-horizon properties differ from the parent-material properties. Ten-point intervals are used in the quantification, and 5-point values can be used for properties with intermediate

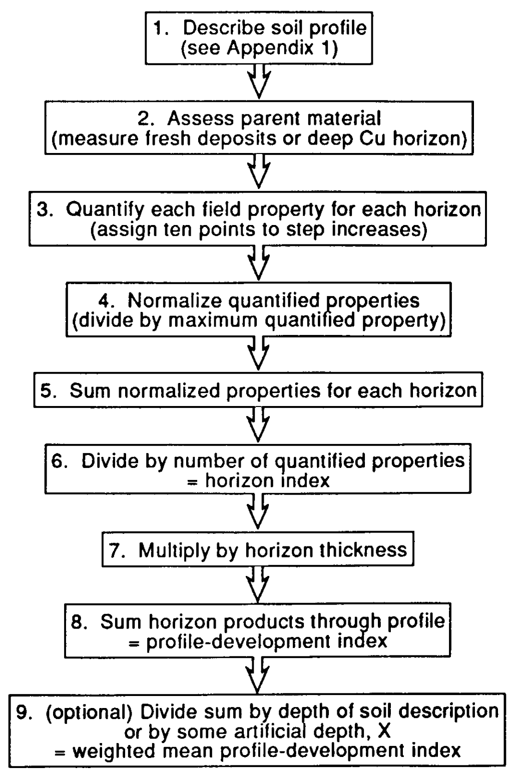


Figure 2-1. Flow diagram for deriving the profile-development index (modified from Harden, 1982; Harden and Taylor, 1983).

Table 2-1.—Quantification and normalization of soil properties for calculating PDI
[Modified from procedures outlined by Harden (1982), Harden and Taylor (1983), and Taylor (1988). See Taylor (1988) for current values (Part B)]

PART A QUANTIFICATION OF SOIL PROPERTIES					PART B NORMALIZATION OF QUANTIFIED PROPERTIES	
SOIL PROPERTY			EXAMPLE Horizon Property	EXAMPLE Parent Material Property	CALCULATION	EXAMPLE
Rubification	10 pts/increase in *hue* redness *5Y* ⇒ *2.5Y* ⇒ *10YR* ⇒ *7.5YR* ⇒ *5YR* ⇒ *2.5YR* ⇒ *10R* ⇒ *5R* PM HP 10 pts/increase in *chroma* HP(d) 0 ⇒ 1 ⇒ 2 ⇒ 3 ⇒ 4 ⇒ 5 ⇒ 6 ⇒ 7 ⇒ 8 PM HP(m)	Note: For multiple colors, record and calculate each.[1] $\chi_r = 10[(\text{hue } \Delta\chi_0) + (\text{chroma } \Delta\chi_0)]_{dry} + 10[(\text{hue } \Delta\chi_0) + (\text{chroma } \Delta\chi_0)]_{moist}$	*7.5YR* 5/4 (d) *7.5YR* 4/5 (m) $\chi_r = 10(1+2)_{dry} + 10(1+3)_{moist}$ $\chi_r = 70$	*10YR* 7/2 (d) *10YR* 6/2 (m)	Divide by current maximum (190) $\chi_{rn} = \chi_r \div 190$	$\chi_{rn} = 70 \div 190$ $\chi_{rn} = 0.37$
Color Paling	10 pts/decrease in *hue* redness *5Y* ⇐ *2.5Y* ⇐ *10YR* ⇐ *7.5YR* ⇐ *5YR* ⇐ *2.5YR* ⇐ *10R* ⇐ *5R* HP PM 10 pts/decrease in *chroma* HP(d) PM(d) 0 ⇐ 1 ⇐ 2 ⇐ 3 ⇐ 4 ⇐ 5 ⇐ 6 ⇐ 7 ⇐ 8 HP PM(m)	$\chi_{cp} = 10[(\text{hue } \Delta\chi_0) + (\text{chroma } \Delta\chi_0)]_{dry} + 10[(\text{hue } \Delta\chi_0) + (\text{chroma } \Delta\chi_0)]_{moist}$	*2.5Y* 6/2 (d) *2.5Y* 6/4 (m) $\chi_{cp} = 10(1+2)_{dry} + 10(1+1)_{moist}$ $\chi_{cp} = 50$	*10YR* 4/4 (d) *10YR* 4/5 (m)	Divide by current maximum (60) $\chi_{cpn} = \chi_{cp} \div 60$	$\chi_{cpn} = 50 \div 60$ $\chi_{cpn} = 0.83$
Melanization	for horizons with upper boundary at ≤100 cm: 10 pts/decrease in *value* HP(d) PM(d) 0 ⇐ 1 ⇐ 2 ⇐ 3 ⇐ 4 ⇐ 5 ⇐ 6 ⇐ 7 ⇐ 8 ⇐ 9 ⇐ 10 HP(m) PM(m)	$\chi_m = 10(\text{value } \Delta\chi_0)_{dry} + 10(\text{value } \Delta\chi_0)_{moist}$	*7.5YR* 5/4 (d) *7.5YR* 4/5 (m) $\chi_m = 10(2)_{dry} + 10(2)_{moist}$ $\chi_m = 40$	*10YR* 7/2 (d) *10YR* 6/2 (m)	Divide by current maximum (85) $\chi_{mn} = \chi_m \div 85$	$\chi_{mn} = 40 \div 85$ $\chi_{mn} = 0.47$
Color Lightening	10 pts/increase in *value* 0 ⇒ 1 ⇒ 2 ⇒ 3 ⇒ 4 ⇒ 5 ⇒ 6 ⇒ 7 ⇒ 8 ⇒ 9 ⇒ 10 PM HP	$\chi_{cl} = 10(\text{value } \Delta\chi_0)_{dry} + 10(\text{value } \Delta\chi_0)_{moist}$	*10YR* 8/2 (d) *10YR* 8/3 (m) $\chi_{cl} = 10(2)_{dry} + 10(2)_{moist}$ $\chi_{cl} = 40$	*10YR* 6/3 (d) *10YR* 6/4 (m)	Divide by current maximum (80) $\chi_{cln} = \chi_{cl} \div 80$	$\chi_{cln} = 40 \div 80$ $\chi_{cln} = 0.50$
Total Texture class	texture 10 pts/line crossing toward clay on texture triangle 10 pts/increase in *stickiness* so ⇒ ss ⇒ s ⇒ vs PM HP 10 pts/increase in *plasticity* po ⇒ ps ⇒ p ⇒ vp PM HP	$\chi_t = 10[(\text{textural } \Delta\chi_0) + (\text{stickiness } \Delta\chi_0) + (\text{plasticity } \Delta\chi_0)]$	SCL s ps $\chi_t = 10(1 + 2 + 1)$ $\chi_t = 40$	SL so po	Divide by current maximum (90) $\chi_{tn} = \chi_t \div 90$	$\chi_{tn} = 40 \div 90$ $\chi_{tn} = 0.44$
Dry Consistence	10 pts/increase in *hardness* lo ⇒ so ⇒ sh ⇒ h ⇒ vh ⇒ eh PM HP	$\chi_{dc} = 10(\text{dry consistence } \Delta\chi_0)$	vh $\chi_{dc} = 10(4)$ $\chi_{dc} = 40$	lo	Divide by 2 x current maximum (2 x 50 = 100) $\chi_{dcn} = \chi_{dc} \div 100$	$\chi_{dcn} = 40 \div 100$ $\chi_{dcn} = 0.40$
Moist Consistence	10 pts/increase in *firmness* lo ⇒ vfr ⇒ fr ⇒ fi ⇒ vfi ⇒ efi PM HP	$\chi_{mc} = 10(\text{moist consistence } \Delta\chi_0)$	fi $\chi_{mc} = 10(3)$ $\chi_{mc} = 30$	lo	Divide by 2 x current maximum (2 x 50 = 100) $\chi_{mcn} = \chi_{mc} \div 100$	$\chi_{mcn} = 30 \div 100$ $\chi_{mcn} = 0.30$

Table 2-1.—Continued

<table>
<tr><th colspan="3">PART A
QUANTIFICATION OF SOIL PROPERTIES</th><th colspan="2">PART B
NORMALIZATION OF QUANTIFIED PROPERTIES</th></tr>
<tr><th colspan="2">SOIL PROPERTY</th><th>EXAMPLE
Horizon Property — Parent Material Property</th><th>CALCULATION</th><th>EXAMPLE</th></tr>
<tr><td>Clay Films[2]
class
<table><tr><td>points</td><td>10</td><td>20</td><td>30</td><td>40</td></tr><tr><td>amount</td><td>vl</td><td>1</td><td>2</td><td>3</td></tr><tr><td>distinctness</td><td>f</td><td>d</td><td>p</td><td></td></tr><tr><td>location</td><td>po
co</td><td>br
pf</td><td></td><td></td></tr></table></td><td>Note: If classes are equal, choose 1° class with greatest abundance.

χ_{cf} = [(abundance + distinctness + location) of 1° class + 1/2 (abundance of 2° class)</td><td>3fbr — no films
1ppf, 2dbr

$\chi_{cf} = (40 + 10 + 20) + 1/2(20 + 30)$
$\chi_{cf} = 95$</td><td>Subtract 20 pts from all $\chi_{cf} > 0$ [3]
Divide by current maximum (130)

$\chi_{cfn} = (\chi_{cf} - 20) + 130 \geq 0$</td><td>$\chi_{cfn} = (95-20) + 130$
$\chi_{cfn} = 0.58$</td></tr>
<tr><td>Structure[4]
class
<table><tr><td>points</td><td>5</td><td>10</td><td>20</td><td>30</td></tr><tr><td>grade</td><td></td><td>1</td><td>2</td><td>3</td></tr><tr><td>type</td><td>pl</td><td>gr
sbk
abk</td><td>pr</td><td>col</td></tr></table></td><td>χ_s = [(grade + type) of 1°] + 1/2 [(grade + type) of 2°]</td><td>1co sbk — structureless

$\chi_s = 10(1) + 10(1)$
$\chi_s = 20$</td><td>Divide by maximum primary possible (60)

$\chi_{sn} = \chi_s + 60$</td><td>$\chi_{sn} = 20 + 60$
$\chi_{sn} = 0.33$</td></tr>
<tr><td>pH[5] — Difference between pH of horizon and parent material.</td><td>$\chi_{pH} = pH\ \Delta\chi_0$</td><td>7.2 — 7.9
$\chi_{pH} = 7.9 - 7.2$
$\chi_{pH} = 0.7$</td><td>Divide by decrease of 3.5 or increase of 1.5
$\chi_{pHn} = \chi_{pH} + 3.5 \geq 0$</td><td>$\chi_{pHn} = 0.7 + 3.5$
$\chi_{pHn} = 0.20$</td></tr>
<tr><td>Carbonate Morphology[6]
Use carbonate stage as multiplier (disseminated, 0.5; stage I to VI, 1 to 6).</td><td>$\chi_{cm} = (\chi_{cp} + \chi_{cl})$ (carbonate stage)</td><td>Using above color paling and color lightening, and stage II carbonate morphology
$\chi_{cm} = (50 + 40) \times 2$
$\chi_{cm} = 180$</td><td>Divide by current maximum (240)

$\chi_{cmn} = \chi_{cm} + 240$</td><td>$\chi_{cmn} = 180 + 240$
$\chi_{cmn} = 0.75$</td></tr>
<tr><td>Color Mottling[7]
10 pts/increase in abundance
none ⇒ few ⇒ common ⇒ many
PM — HP

10 pts/increase in degree of contrast
none ⇒ faint ⇒ distinct ⇒ prominent
PM — HP</td><td>χ_{mo} = abundance $\Delta\chi_0$ + contrast $\Delta\chi_0$</td><td>common — none
2.5YR 4/6 (m) — 2.5Y 7/3 (m)

$\chi_{mo} = 10(2) + 10(3)$
$\chi_{mo} = 50$</td><td>Divide by current maximum (60)

$\chi_{mon} = \chi_{mo} \div 60$</td><td>$\chi_{mon} = 50 + 60$
$\chi_{mon} = 0.83$</td></tr>
<tr><td>Clast Weathering[8]
10 pts/increase in weathering stage
none ⇒ slight ⇒ moderate ⇒ high ⇒ extreme
PM — HP</td><td>χ_{cw} = 10 (weathering stage $\Delta\chi_0$)</td><td>moderate — none
$\chi_{cw} = 10\ (2)$
$\chi_{cw} = 20$</td><td>Divide by current maximum (40)
$\chi_{cwn} = \chi_{cw} + 40$</td><td>$\chi_{cwn} = 20 + 40$
$\chi_{cwn} = 0.50$</td></tr>
</table>

1 See Taylor (1988) for details of calculating multiple colors.
2 For this property and structure property, 1° denotes primary and 2° secondary. The new terms amount and distinctness replace frequency and thickness. Workers might consider a change in location points: br = 10; co, cobr, po = 20; pf = 30. We believe this is a more common progression with time.
3 Harden (1982) subtracts 20 pts; we do not, and instead compare the horizon to the parent material, and record only positive values.
4 Workers might consider a change in type points: gr, sbk = 10; abk, pr, pl = 20; col = 30.
5 The original intent here was to assign points for a lowering of pH. However, in some places the pH will increase with time as pedogenic carbonate accumulates. If this is done, use a maximum pH increase of 1.5 (maximum pH of 8.5).
6 Stages from Bachman and Machette (1977), Gile and Grossman (1979), and Machette (1985a). Modified by Sowers and others (1988), and Taylor (1988).
7 From Knuepfer (1988).
8 From Knuepfer (1988). The stages are slightly weathered (minor spalling and cracking), moderately weathered (easily broken by hammer), highly weathered (disintegrated by hand), and extremely weathered (so completely weathered that only outline of clast remains). Workers can alter this for different kinds of rocks and progressive weathering stages.

values. It should be emphasized that these point assignments for various degrees of development of a property are somewhat arbitrary and a soil property that is assigned 20 points does not necessarily mean the property is twice as developed as one that is assigned 10 points. The number of properties is unlimited, with 13 given in Table 2-1. In each case, the direction of point increase is the direction of the progression of that property with time. Several rules apply, however. If rubification proceeds with time, and color paling does not occur within the soil sequence of interest, the latter should not be included as a zero property but should be excluded from the index. The same goes for the melanization-color lightening pair. However, if clay films occur in older soils but not in younger soils, they should be entered as zero properties in the younger soils.

4. The properties are then normalized to what is called the "current maximum" (see Table 2-1, part B). This results in a number between 0 (no development) and 1 (maximum development) for that property. However, in places the horizon value might exceed that of the current maximum (*i.e.*, 1.25), and rather than redefine the current maximum, we proceed with the calculation, using the normalized value of 1.25. The system is very flexible and allows for alterations as given in Table 2-1, as well as the introduction of new properties.
5. Sum the normalized values for properties for each horizon.
6. Divide the sum (in 5) by the total number of properties for that horizon. This value is called the horizon index, which varies between 0 and 1 unless the soil horizon you are describing exceeds the current maxima. In the case where a horizon has a zero property, it should be included in the total number of properties. As in the example mentioned earlier, young soils may not have any clay films yet older soils do. The property value therefore is zero in the young soils, and it is included in the number of total properties investigated.
7. Multiply by horizon thickness in centimeters in order to weight the horizon properties for the proportion of the total profile that the horizon represents.
8. Sum the horizon products for the profile. This is the profile-development index (PDI) of most workers. The value is open ended at the high end.
9. Divide sum (in 8) by depth of the described soil or by some prescribed (artificial, in places) depth. This produces a weighted mean PDI, with values between 0 (no development) and 1 (maximum development). The depth used in this calculation is important. Most soils are described to variable depths, usually to the parent material; in particular, young soils are described to shallow depths and old soils to great depths. However, if the Cu horizon begins at 20 cm in a young soil and at 120 cm in an old soil, the best comparison between the two is to dig both to 120 cm, but this is seldom done. The resulting PDI's for the young soil can differ significantly depending on the total soil depth considered. If the PDI for the young soil is weighted to the same depth as the old soil, the inclusion of 100 cm of zero properties (Cu horizon) for the younger soil will result in a smaller PDI for the young soil than if its properties are weighted to the maximum depth of the hole. The result is a greater difference in PDI's between young and old soils depending on the method used. It may be preferable to compare soils to an equivalent apparent described thickness (termed artificial in Fig. 2-1) by increasing the thickness of the lowest horizon in all but the thickest soil in the local sequence. The problem with this manipulation is the assumption that the properties of the lowest horizon in all profiles continues to that uniform depth, and this may not be true for all soils. J.W. Harden (written commun., 1990) does not endorse these manipulations and prefers these be termed optional. Table 2-1 gives examples of the calculations.

PLOTTING AND COMPARING SOIL-INDEX DATA

There are many ways to plot either the horizon- or profile-development index as illustrated in the following examples.

1. Plot individual properties with depth to see changes that take place over the same time range in different environments (Fig. 2-2).
2. Plot profile-property data *versus* age for soils in different environments to show the variation in development with environment (Fig. 2-3).
3. Plot the horizon-development-index data with depth to portray changes in soils with time (Fig. 2-4).
4. Plot PDI *versus* estimated age (Fig. 2-5) to portray soil development as a function of deposit age. The data are best portrayed as semi-log or log-log plots owing to the great range in soil age that is typically encountered in soil chronosequence studies (*i.e.*, several thousand years to a million years).
5. Finally, one can compare PDI with time for different environments (Fig. 2-6). Various manipulations can be made; Harden and Taylor (1983) preferred using the four "best" properties in each environment (not the same properties in each environment) rather than all eight of the properties for which they collected data. When the four best properties were used, the data from soils formed in greatly different environments plotted within a tighter envelope (curve) than when all eight properties were used.

Many applications can be made of the PDI data. However, of primary concern to geologists is the estimation of the ages of surficial deposits. Several examples of this use will be made in the following sections.

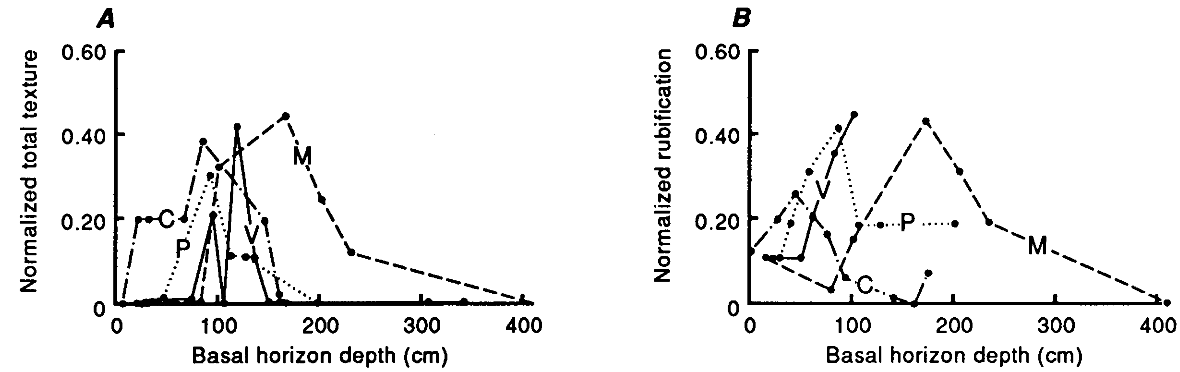


Figure 2-2. Depth plots of two field properties of soils that are approximately 40-120 ka (from Harden and Taylor, 1983). ***A***. Normalized total texture plotted against basal horizon depth. ***B***. Normalized rubification plotted against basal horizon depth. Locations are V-Ventura, California; M-Merced, California; C-Las Cruces, New Mexico; P-Pennsylvania.

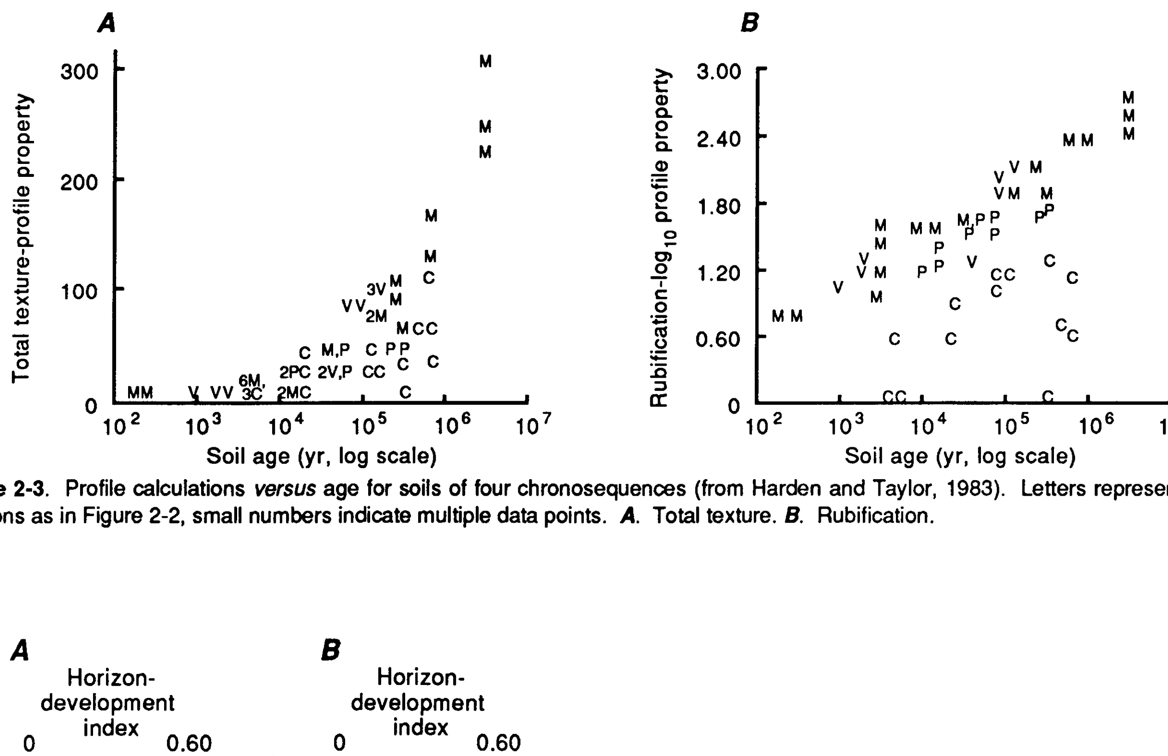


Figure 2-3. Profile calculations *versus* age for soils of four chronosequences (from Harden and Taylor, 1983). Letters represent same locations as in Figure 2-2, small numbers indicate multiple data points. ***A***. Total texture. ***B***. Rubification.

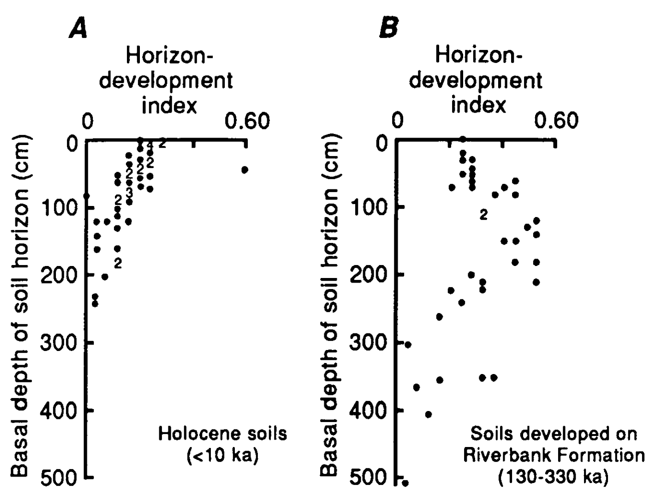


Figure 2-4. Horizon-development index *versus* depth to base of horizon for two different age soils near Merced, California (from Harden, 1982). Small numbers indicate multiple data points. ***A***. Holocene soils. ***B***. Soils on Riverbank Formation, estimated at 130-330 ka.

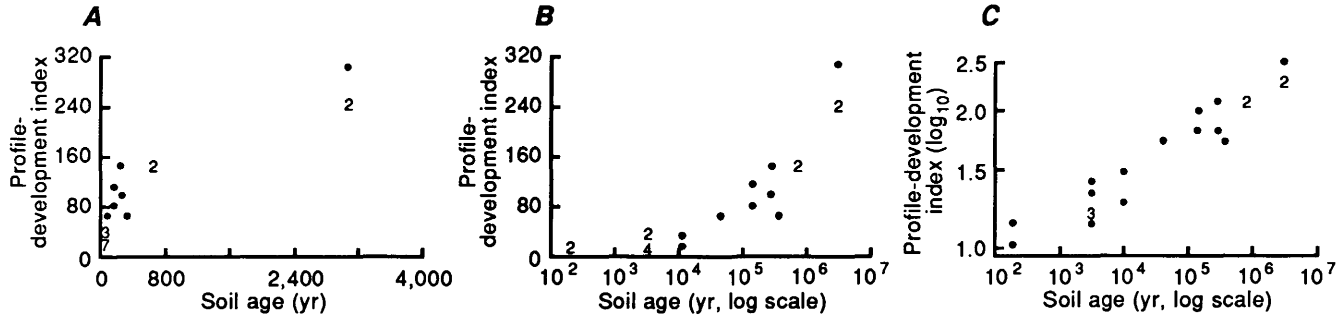


Figure 2-5. Profile-development index (PDI) *versus* age for soils near Merced, California (from Harden, 1982). Small numbers indicate multiple data points. The PDI *versus* soil age are in (**A**) yr, (**B**) $\log_{10}$ yr; (**C**) PDI ($\log_{10}$) *versus* soil age in $\log_{10}$ yr.

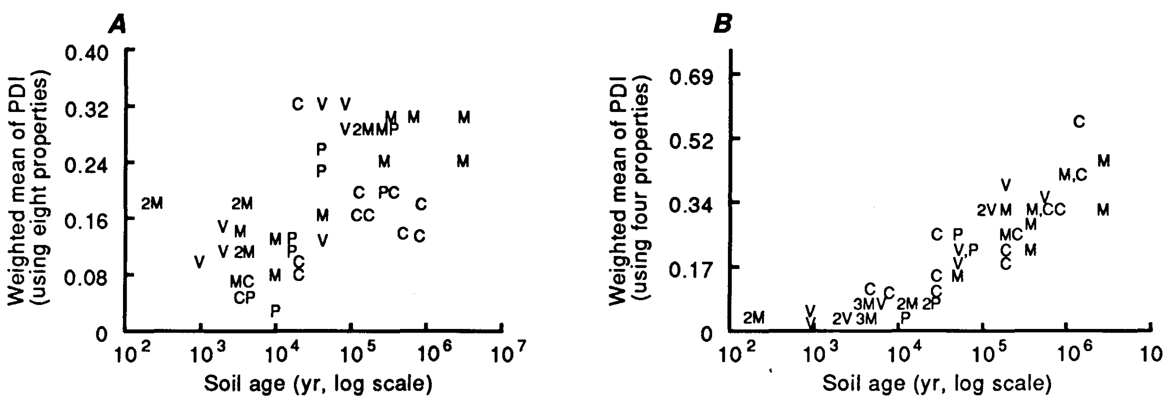


Figure 2-6. PDI calculations for soils of four chronosequences in different environments *versus* age (from Harden and Taylor, 1983). Letters represent same locations as in Figure 2-2. ***A***. PDI calculated using eight soil properties. ***B***. PDI calculated using four soil properties.

CHAPTER III
PROCESSES AND FACTORS OF SOIL FORMATION WITH EMPHASIS ON THE PARENT-MATERIAL FACTOR

PROCESSES OF SOIL FORMATION

Many processes act together to form a soil. The formation of a soil profile was viewed by Simonson (1978) as the combined effect of additions to the ground surface, transformations within the soil, vertical transfers (up or down) within the soil, and removals from the soil (Fig. 3-1). For any one soil, the relative importance of these processes varies, and the result is the wide variety of soil profiles seen in the landscape. The main additions to most soils are organic matter from surface vegetation and its contained elements, ions and solid particles introduced with rainfall, and particles carried by the wind. Transformations include the multitude of organic compounds that form during decomposition of organic matter, the weathering of primary minerals, and the formation of secondary minerals (*i.e.*, clay minerals), and other products such as amorphous and crystalline iron compounds. Transfers generally involve the movement of ions and suspended substances in moving soil water. Soluble substances move with the percolating water unless changing chemical conditions or dehydration causes them to precipitate from solution. In places where capillary rise of ground water is important, ions can be transferred upward and be precipitated high in the soil profile. Ions can also move upward through plants and be returned to the surface with plant litter. Soil-dwelling fauna can actively move solid particles in any direction. In addition, when water moves through the profile, it removes substances still in solution; these substances then become part of the dissolved constituents of the ground or surface waters.

It should be pointed out that it is difficult to determine the magnitude and importance of individual soil processes because so few actual measurements can be or have been made. Consider, for example, clay translocation from the A to the B horizon. Translocation probably is so slow that measurements taken over a year or two, or even a decade may not be representative of the long-term processes leading to soil formation. In addition, laboratory simulations may or may not be good approximations of what happens in nature. In lieu of these approaches, we commonly infer the processes from combined field and laboratory data. Interpretation is confounded because, although the soil-formation process is commonly tied to particular climates, climates have varied during the formation of many Quaternary soils; hence, what one might infer to be a contemporary process actually could have been active only at some time in the past. One should not always expect variations in processes or their rates with climatic change, for it might be that neither of these changed markedly during the time of soil formation. However, climatic change seems fairly well documented in the Quaternary geologic record, and we should attempt to see how well soil-process models and rates fit with the degree, kind, and timing of climatic change.

Translocation of Clay-Size Particles

The distribution of clay-size particles in many moderately to strongly developed soils is marked by relatively low clay contents in the A and C horizons with the maximum amount in the Bt horizon, generally in its upper part (the clay bulge). This distribution can be accounted for in three ways: (1) Dissolved constituents derived by weathering higher in the profile (A horizon) move downward in solution with the percolating water and precipitate as clay minerals in the B horizon. (2) The clays form *in situ* from mineral weathering in the B horizon. (3) The clays move as particles in suspension in the downward-percolating water to accumulate in the B horizon because of flocculation or constrictions in the pores through which the water moves, or because the base of the B horizon marks the lower limit of most water movement (McKeague and St. Arnaud, 1969). There is no doubt that in most soils, clays in the B horizon form in all

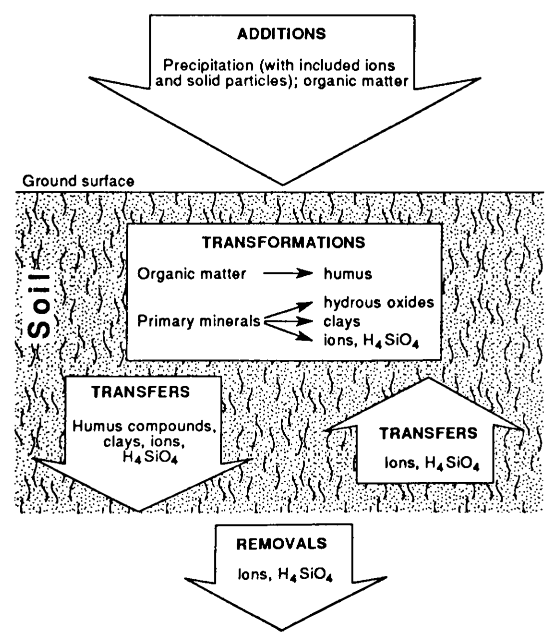


Figure 3-1. Flow chart of major processes in soil-profile development (from Simonson, 1978). Surface erosion might be added, but erosion rates must be less than soil-formation rates for profiles to form.

three ways, but the relative importance of each may vary from soil to soil, region to region, and climate to climate. It may be possible to differentiate between translocated clay particles and those formed *in situ*; however, there are no known criteria for identifying those clays that have precipitated from downward-moving solutions.

Ratios of clay fractions, such as fine clay to total clay, can be used to indicate clay translocation (Walker and Hutka, 1979). The hypothesis is that the finer clay sizes are more mobile and, thus, will move more readily from the A to the B horizon, or downward within the B horizon, thereby resulting in a higher ratio of fine clay to coarse clay in the B horizon. In most cases, however, the interpretation may not be so clear, because some ratios can be explained by parent-material layering or because the fine clay could be clay minerals newly formed from precipitation from solution.

Clay films that coat ped surfaces or line voids can be seen in both field and thin-section studies; these films generally are taken as evidence for clay-particle translocation. This translocation can be verified by comparing detailed chemical analyses of the clay in the horizon with those in the film; the two usually differ because the film has been emplaced by a more recent downward movement. Therefore, the film can have properties that more closely resemble those of the A horizon than of the B horizon. Thin-section analysis indicates that films of oriented, fairly pure clay along voids and ped surfaces that have sharp boundaries with the soil mass probably result from translocation (Kemp, 1985). Care must be taken, however, to eliminate the possibility that the oriented clay particles are caused by stresses induced during the shrinkage and swelling of soils during dry and wet cycles, respectively.

One problem with using clay films as the only basis of demonstrating clay illuviation is that they are sometimes destroyed as soon as they form, or later. Nettleton and others (1969) reported clay-film destruction due to shrinkage and swelling in soils that formed in the desert and Mediterranean climates of the southwestern United States. They found, for example, that soils with a low shrink-swell potential and less than 40% clay content can retain clay films and that soils with a high shrink-swell potential and higher than 40% clay content cannot retain films. Thus, many soils may have translocated clay, but the thin-section evidence for it is destroyed as the films become incorporated in the matrix of the B horizon. In the field, some clay-enriched Bt horizons have what appears to be clay films, but instead, these may be incipient slickensides on ped faces. In places, these incipient slickensides have a subtle sheen. Gile and others (1981) noted that in desert regions, clay films are most stable on the surfaces of sand and pebbles, but they can be destroyed by either the accumulation of $CaCO_3$ in the soil or by bioturbation (mixing by animals).

Flocculation and dispersion can decrease or increase, respectively, clay movement in soils (McKeague and St. Arnaud, 1969). Calcium-rich clays commonly are flocculated and, thus, do not migrate easily. Soils enriched in $CaCO_3$, for example, show little evidence of clay migration; most migration takes place after the carbonate has been leached from the soil and some of the Ca^{2+} in the exchange complex has been replaced by other ions. In contrast, clays with appreciable amounts of exchangeable Na^+ remain dispersed, and clay movement can be rapid. The dispersive effects of Na^+ is such that in some basins of the deserts of the Western United States, Na^+-influenced Bt horizons can form in about one-half the time it takes Bt horizons to form elsewhere (Alexander and Nettleton, 1977; Peterson, 1980). The same effect is seen in some coastal environments (Muhs, 1982a). With still higher concentrations of Na^+, however, flocculation of clays can result.

Conditions for clay migration can vary with depth in the profile, or with seasonal changes. Quite commonly, the electrolyte content will increase with depth because the upper parts of the profile have more water moving through them. Thus, clays could be dispersed near the surface, but be flocculated at depth. In soils with a high clay content, high shrink-swell potential, and distinct dry and wet periods (*i.e.*, Mediterranean climate), clay movement may take place mainly at the onset of the wet season, when an open prismatic soil structure with wide cracks reaches the surface and provides avenues for rapid translocation of clay.

Still another way in which clays can be translocated in a soil is by suspension initially in rainwater. Holliday (1988) mentioned this as the mechanism that formed the Bt horizons in as short a period as 450 yr in the Lubbock, Texas, area. Here, it is not uncommon for a dust storm to precede a thunderstorm, and the subsequent rain commonly comes through the dust, producing a "mud rain." Such occurrences are not mentioned in the literature dealing with translocated clay, but they are common in the atmospheric and meteorology literature; they may be an important soil-forming process in certain regions. One appealing aspect of this process is that it does not require that clays already present in the soil be brought into suspension before translocation by slowly moving soil waters. A similar mechanism might also be responsible for the movement of clays in calcareous materials. It was argued above that such movement would not be expected, but it has been demonstrated in an artificial field experiment (Goss and others, 1973).

Origin of $CaCO_3$-Rich Horizons and Those with More Soluble Salts

Soils in semiarid and arid regions commonly have carbonate-rich horizons at some depth below the surface, or if the climate is dry enough or the surface erosion intensive enough, these horizons may extend to the surface. Although several origins have been presented for some of these horizons (Goudie, 1973; Reeves, 1976), our concern here is with $CaCO_3$-rich horizons of pedogenic origin. Some of these horizons are the caliche and calcrete of present and past geologic literature. Pedologists call these soil accumulations Bk and K horizons, and these are preferable to the general terms caliche or calcrete. Both calcium and magnesium carbonates are present in these soils with calcium being the dominant carbonate.

The origin of carbonate horizons involves carbonate-bicarbonate equilibria, as shown by the following reactions:

$$\begin{array}{cccc} & \underset{(g)}{CO_2} + \underset{(l)}{H_2O} & & \\ & \updownarrow & & \\ \underset{(c)}{CaCO_3} + & \underset{(aq)}{H_2CO_3} & \Leftrightarrow \underset{(aq)}{Ca^{2+}} & + \underset{(aq)}{2HCO_3^-} \end{array}$$

An increase in CO_2 content in the soil air, or a decrease in pH in soil water will drive the reaction to the right; carbonate will dissolve and move as Ca^{2+} and HCO_3^- with the soil water. Dissolution is also favored by increasing the amount of water moving through the soil, as long as the water is not already saturated with respect to $CaCO_3$. Precipitation of carbonate occurs under conditions that drive the reaction to the left, that is, a lowering of CO_2 pressure, a rise in pH, an increase in ion concentration to the point where saturation is reached and precipitation takes place, or evapotranspiration of the soil moisture.

All of the above conditions are found in soils in which $CaCO_3$ has accumulated. Carbon dioxide partial pressures in soil air are 10 to more than 100 times greater than that found in the atmosphere (Brook and others, 1983; Quade and others, 1989); this change in partial pressure decreases the pH, which in turn increases the solubility of $CaCO_3$. The partial pressure of CO_2 is high as a result of CO_2 produced by root and microorganism respiration and organic-matter decomposition. Thus, one would expect the highest CO_2 partial pressure to be associated with the A horizon, with values diminishing down to or beyond the base of the root zone. The amount of water leaching through the soil also is greater near the surface than at depth, so as the water moves vertically through the soil, the Ca^{2+} and HCO_3^- content might increase to the point of saturation after which further dissolution of $CaCO_3$ is not possible. Combining the effects of high CO_2 partial pressure and downward-percolating water, we might visualize the formation of a horizon enriched in $CaCO_3$ as follows.

1. In the upper parts of the soil, Ca^{2+} may already be present or may be derived by weathering of calcium-bearing minerals. Due to plant growth and biological activity, CO_2 partial pressure is high and forms HCO_3^- upon contact with water.
2. Water leaching through the profile can carry the Ca^{2+} and HCO_3^- downward in the profile.
3. Precipitation as a $CaCO_3$-rich horizon would take place by a combination of decreasing CO_2 partial pressure below the zone of rooting and major biological activity and the progressive increase in concentration in Ca^{2+} and HCO_3^- in the soil solution with depth as (a) the water percolates downward and (b) water is lost by evapotranspiration.

The position of the $CaCO_3$-bearing horizon is therefore, related to depth of leaching, which, in turn, is related to the climate (Goudie, 1973; McFadden and Tinsley, 1985; Mayer and others, 1988).

Several stages in the buildup of carbonate horizons are recognized (Gile and others, 1981; Machette, 1985a). In gravelly material, the first stage is the appearance of carbonate coatings on the undersides of gravel particles; in nongravelly material, the first stage is the occurrence of thin filaments (Fig. 1-2; Appendix, Tables A-6 and A-7). The undersides of gravel are favored sites, initially, because downward-moving water tends to collect there. With time, in both types of parent materials, the horizon is increasingly impregnated by carbonate deposition as segregated masses in the matrix and as coatings on solids until the voids become plugged and water percolation through the horizon is greatly restricted. At this point, water tends to collect periodically over the nearly plugged horizon; the resulting solution and reprecipitation produces the laminated part of the upper K horizon. At this point in the development, the horizon builds upward and so is younger in that direction. Sowers and others (1988), however, reported that this model could be oversimplified, because at Kyle Canyon, Nevada, there are several plugged layers of different ages that are crosscut by vertical cracks filled with younger carbonate. During the buildup of a K horizon, the volume of pedogenic carbonate eventually exceeds the original volume of the pores. In most cases, this is attributed to the crystallization of carbonate that forces the mineral grains and gravel clasts apart (Watts, 1978; Machette, 1985a), but there is also evidence of carbonate replacing silicate minerals (Reheis, 1988). The buildup of $CaCO_3$ in the diagnostic stages (see Appendix, Tables A-6 and A-7) is more rapid in gravel than in nongravelly material because gravel has a lower initial porosity and lower surface area to volume ratio. Finally, progressively higher stages are usually attained in the older soils of a chronosequence; in any one soil profile, however, the maximum stage is usually near the top of the carbonate-enriched K horizon, and this grades to progressively lesser stages with depth.

Several other morphological features are present in indurated K horizons that usually have morphologies of stages IV, V, and VI. One is ooids, which are sand-sized round particles in which a nucleus is surrounded by one or more concentric layers of fine-grained calcite grains (micrite). The origin of ooids is not clear, but those layers studied by Hay and Reeder (1978) seem to result from carbonate replacing clay coats on sand grains. Another common feature is pisoliths, which are subangular to spheroidal bodies, 0.5 cm to more than 10 cm across, surrounded by many thin layers of micrite. The nuclei of pisoliths can be any material, from rock fragments to pieces of broken and rotated indurated K-horizon material, some of which display internal laminar layers and pisoliths from previous episodes of pedogenesis. Both of these features usually are set into massive carbonate cement (Fig. 3-2*A*). The most advanced stage of carbonate morphology (VI) is characterized by pisolith formation, multiple generations of brecciation of the upper parts of calcretes, and pervasive recementation (Fig. 3-2*B*). The calcretes of the Pliocene Mormon Mesa surface in southern Nevada (Gardner, 1972; Machette, 1985a) and the upper Miocene Ogallala surface in eastern New Mexico and western Texas (Bachman and Machette, 1977; Machette, 1985a) are two of the oldest and morphologically complex soils preserved in the Western United States.

The $CaCO_3$ of the carbonate horizons may come from several sources. The Ca^{2+} released by weathering of primary minerals could combine with HCO_3^- deeper in the profile, and if so, there should be a close relation between the CaO content of the parent material, the amount of weathering in the upper part of the soil to release Ca^{2+}, and the amount of $CaCO_3$ in the Bk or K horizon. However, for most K horizons in the Western United States, derivation of the carbonate by parent-material weathering is highly unlikely, owing to semiarid climatic conditions. Gardner (1972) has calculated that 37-90 m of material would have to be weathered to release the amount of Ca^{2+} present in the K horizon of Mormon Mesa, Nevada. Not only does the lack of mineral weathering negate such an origin, but geomorphologically, it is highly unlikely. For example, if the K horizon forms by parent-material weathering, tens of meters of weathered material would have to be removed from the mesa; however, the carbonate forms a horizon several meters thick, immediately below flat geomorphic surfaces. In many places, an external source of Ca^{2+} seems most likely because many soil parent materials are noncalcareous. Detailed study in the Las Cruces region of New Mexico indicates that the atmosphere is an important source for Ca^{2+}-rich dust and $CaCO_3$ (Gile and others, 1981). Dust-trap data at Las Cruces suggest that the dust contains less than 5% carbonate, and is added to the soil at a rate of about 0.2-0.4 x 10^{-4} g/cm^2/yr. However, the amount of dissolved Ca^{2+} in yearly precipitation is such that it produces perhaps three times that amount of carbonate, giving total annual carbonate production of about 2 x 10^{-4} g/cm^2/yr. This can only be an estimate because, although the Ca^{2+} in precipitation probably is car-

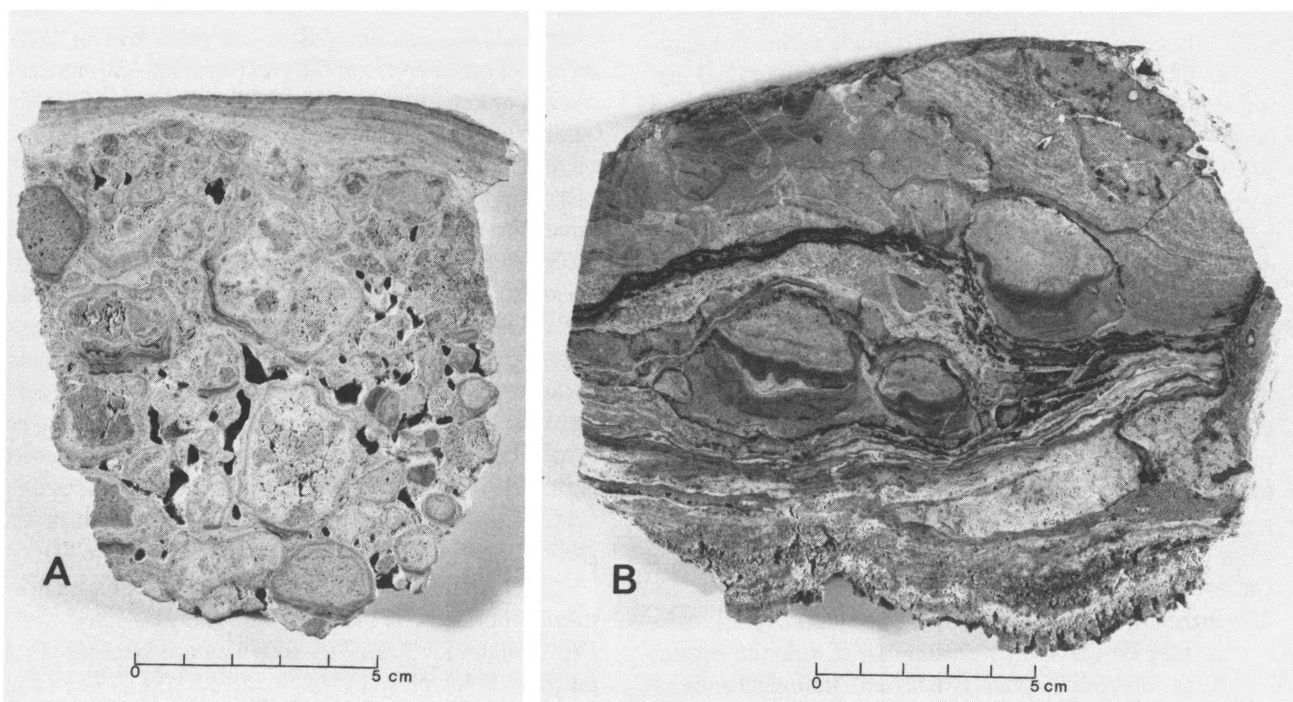


Figure 3-2. Polished sections of Km horizons with advanced stages of carbonate morphology (from Machette, 1985a). ***A***. Incipient stage V morphology found in the upper part of the pedogenic calcrete on the early Pleistocene upper La Mesa surface, Las Cruces, New Mexico. Sample is almost entirely $CaCO_3$, with pisoliths (p; several outlined) composed of porous but densely cemented calcrete set in a matrix of massive carbonate (m), and the top marked by 1-cm-thick layer of laminar carbonate (l). ***B***. Stage VI morphology typically found in the upper part of the caprock (pedogenic calcrete) of the Ogallala Formation (upper Miocene) in eastern New Mexico and western Texas. Sample shows complex internal stratigraphy in which old pieces of brecciated calcrete (b) are recemented by many generations of laminar carbonate (l) in multiple cross-cutting relations.

ried into the soil, the carbonate in dust could be carried to other sites before precipitation events dissolve and move it into the soil. A similar atmospheric origin for most pedogenic carbonate is favored for semiarid to arid regions in the Western United States (Machette, 1985a).

In areas of extreme aridity (precipitation <300 mm/yr), salts more soluble than the carbonates may accumulate at depth in well-drained soils, yet well above the influence of the water table (Dan and Yaalon, 1982; Nettleton and others, 1982). The common salts in arid soils are gypsum ($CaSO_4 \cdot 2H_2O$) and halite (NaCl). Some horizons contain over 90% gypsum, and at such high contents, the original silicate grains are forced aside during gypsum crystallization. These salts accumulate at depths reached by soil waters on an annual basis, or by extreme rainfall events, and if both salt accumulations occur in one profile, the depth to the maximum amount of halite is typically deeper than that for gypsum because halite is more soluble.

The source of the salts is a problem. Some are derived from salts in the weathered rock or the host sediment. A source for sulfur could be its release during the weathering of pyrite. If the parent material is sufficiently low in salts, however, pedogenic salts probably have an atmospheric origin, either as solid matter or as ions and cations in precipitation. In coastal areas, marine aerosols are the likely sources of gypsum in soils (Muhs, 1982a).

It is not uncommon for the horizon of carbonate or gypsum accumulation to form at the surface. This might be due to extreme aridity, derivation from a former shallow water table, or erosion of all overlying horizons. Knowing the present field relations, the geological history of the area, and the mean-annual precipitation might allow one origin to be favored over another. For example, the precipitation in many parts of the Western United States is such that Km horizons should be at a minimum depth of several tens of centimeters; where they are not, erosion of overlying horizons is suspected.

Numerical Dating of Carbonate Horizons

Although numerous attempts have been made to date carbonate horizons, by both radiocarbon and uranium-series methods (see Birkeland, 1984), the results have been mixed. One of the latest attempts at such dating is from Kyle Canyon, Nevada, where there are a series of alluvial fans of different ages (Sowers and others, 1988; Table 3-1). Samples were obtained from pedogenic carbonate rinds formed on

Table 3-1.—Radiometric ages for pedogenic $CaCO_3$ rinds from clasts in soils on the Kyle Canyon fans, southwestern Nevada [Ages are in ka (thousands of years ago). From Sowers and others (1988)]

	INNER RINDS			OUTER RINDS	
SAMPLE	U-Th	$^{14}C(CaCO_3)$	^{14}C(organic)	SAMPLE	$^{14}C(CaCO_3)$
Surface 1 at 1,400 m					
53-283-B	110 +70, -40				
53-284	110 +70, -40				
7A-1 (eroded)	88 ± 12				
10-401-B	>350				
Surface 2 at 1,400-1,670 m					
17-320-A,C,D,E	129 ± 6†				
17-320-B	129 ± 6				
82-322-B	129 ± 6				
22-308,312,313	129 ± 6				
9A-3	129 ± 6	34.1 ± 0.9	18.38 ± 0.34		
11-1	129 ± 6				
9/10	129 ± 6				
Surface 3 at 1,750 m					
				P-1	9.28 ± 0.12
Surface 3 at 1,400 m					
5/6	76 ± 6†	10.64 ± 0.17	4.40 ± 0.06	JT-E	19.46 ± 0.22
5-1	76 ± 6	12.37 ± 0.13		JT-A	12.82 ± 0.13
5-2		15.18 ± 0.22			
Surface 3 at 840 m					
1B-3		11.97 ± 0.11		C-2	16.70 ± 0.17
1B-4	47 ± 20	11.47 ± 0.13		C-3	29.52 ± 0.60
Reworked rinds, Surface 3 at 840 m					
1B-1	75 ± 20	30.80 ± 0.50	11.09 ± 0.10		
1B-2		24.80 ± 0.50	22.50 ± 0.22		
Surface 4 at 2,150 m					
				PP-1	7.97 ± 0.09

† Samples were individually analyzed, then averaged using least-squares fitting.

limestone clasts in the soils, and dates were obtained for the inner rind by the U-series method, and inner and outer rinds by AMS radiocarbon dating of both the inorganic carbon in $CaCO_3$ and the organic carbon. The two methods give widely different results (Table 3-1), and there are even some reversals in ages. Most dates seem to indicate the minimum age of the deposit. In a study in northern Mexico, for example, Slate and others (1991) suggest that the U-series dates might be only one half of the "true" ages of the deposits, as did Machette (1985a) for U-series dates by Ku and others (1979) from carbonate rinds in soils at Vidal Junction, California. Sowers and others (1988) listed these potential sources of error: (1) contamination of pedogenic carbonate with limestone (detrital or from solution), making the date too old; (2) continued solution and reprecipitation of the carbonate, making the date too old; (3) leaching of U from rinds, making the date too young; and (4) exchange of mobile organic carbon between rind and soil, making the date too young. They ended by saying that they could draw no conclusions from these data regarding the true ages of the deposits! Age estimates based on the mass accumulation of carbonate in soils might still provide the best age information (Chapter IV).

FACTORS OF SOIL FORMATION

Five factors commonly used to define the state of the soil system are climate, organisms, topography, parent material, and time (Jenny, 1980). However, other factors may be important locally, such as atmospheric dust of various composition (clay minerals, salts, *etc.*) in aridic regions. The definitions of the "state factors" of Jenny are as follows:

Climate (cl)—The regional climate; because precipitation and temperature are often not interdependent, they can be treated as separate functions.

Organism or biotic factor (o)—Because vegetation is most important, this factor is the potential vegetation; it is approximated by a list of species growing in the surrounding region that could gain access to the site under appropriate conditions. This is a fairly ideal definition and useful for some purposes. However, vegetation type does influence soil formation in important ways (Jenny, 1980; Birkeland, 1984), and should be listed in any study of soils. Past vegetation might also be incorporated into the interpretation, if known. Because climate and vegetation type are intimately interrelated, one should perhaps combine both as a bioclimate factor.

Topography (r)—Included are the shape and slope of the landscape related to the soil, the direction the slope faces (aspect), and the effects of a high water table (mottling, salt accumulation, *etc.*), the latter being common in the lower parts of a landscape.

Parent material or initial state of the system (p)—Included are materials, both weathered and unweathered, from which the soil formed. Parent material could also be a soil in the case where one wishes to study the effect of climatic change on a preexisting soil.

Time (t)—Elapsed time since deposition of material, the exposure of the material at the surface or formation of the slope to which the soil relates; if a study is being made of the effect of climatic change on a preexisting soil, it is the time since the change occurred that is important.

Equations can be formulated to establish the dependence of certain soil properties on the factors. The most widely quoted one is the fundamental equation of Jenny (1941):

$$S \text{ or } s = f(cl, o, r, p, t, \ldots)$$

where S denotes the soil, s denotes any soil property, and the dots after t represent unspecified factors, such as eolian materials (dust, salt, *etc.*) that might be important locally. In this equation, S and s are the dependent variables; the factors are the independent variables in that field sites can be found where the factors vary independently. Although the factors can be dependent variables in some field sites, their real value in a rigorous quantitative factorial treatment is as independent variables.

Jenny rewrote the equation by solving for one factor at a time. To do this, one factor is allowed to vary while the others are held constant. He established the following functions:

$s=f(\underline{cl}, o, r, p, t, \ldots)$ climofunction,
$s=f(\underline{o}, cl, r, p, t, \ldots)$ biofunction,
$s=f(\underline{r}, cl, o, p, t, \ldots)$ topofunction,
$s=f(\underline{p}, cl, o, r, t, \ldots)$ lithofunction, and
$s=f(\underline{t}, cl, o, r, p, \ldots)$ chronofunction.

To solve each function, the first factor (underlined) is allowed to vary while the others remain constant. Therefore, one determines the dependency of one (or more) soil property on a single factor by appropriate statistical methods. This can be extended to include more factors, and eventually it might be possible to rank them on their relative importance to that soil property or properties.

There are two ways in which a factor can be considered constant: (1) if the range in the state factor is quite small and (2) if variation in the "state factor" is large, yet has a negligible effect on the soil property. Figure 3-3 illustrates the case of large variation. The functional relation between a soil property and *factor a* is to be determined. In the same sampling area, *factor b* also varies, so its relation to the soil property must also be established. The plot (Fig. 3-3) indicates a close dependence of the property on both factors between 0 and x on the horizontal axis, and a simple functional relation cannot be established. Between x and y, however, ds/dF_b (the slope of the curve) approaches zero, and, thus, variation in *factor b* has little effect on the soil property. Therefore, in the equation it can be considered a constant. Variation in s, therefore, between x and y is ascribed to variation in *factor a* in this simplified model. Beyond y, both factors can be considered constants because the slopes of the

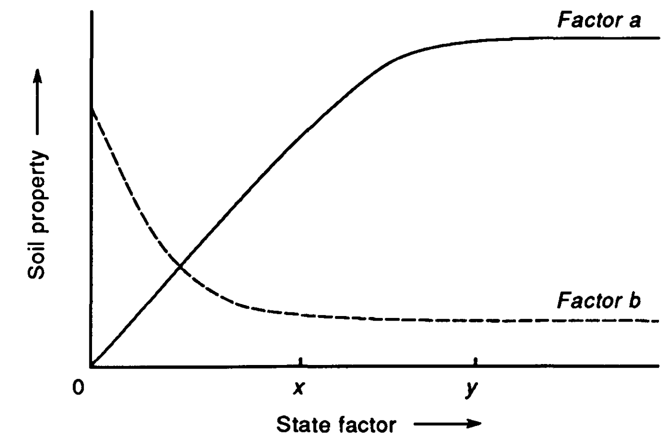


Figure 3-3. Hypothetical variation in one soil property with variation in two "state factors" (*a* and *b*). The vertical scale is not necessarily the same for both plots. On the horizontal axis, x and y are arbitrary values for the state factors.

functions are close to zero. To establish the dependence of *s* on *factor a* between 0 and *x*, sites would have to be chosen in which variation in *factor b* is small.

A major problem with the factorial approach is that many soils that began forming in the Pleistocene probably have formed under a fluctuating bioclimatic regime; for example, under alternating interpluvial and pluvial climates. These effects are readily seen in some soils but not in others. Such variations might produce steps in curves of some soil property *versus* time, especially as we learn more of past climatic change and the ages of soils.

THE PARENT-MATERIAL FACTOR

Parent material is treated briefly here, because it exerts several important controls on soil formation (Birkeland, 1984).

1. Rocks of different mineralogies weather at different rates. Therefore, if pedogenic clays or carbonate are related to weathering rather than to atmospheric sources, their accumulation is controlled by the rates of weathering reactions. Consider a limestone parent material in a semiarid environment. Clays in the soil would be those from the insoluble residue as the limestone dissolves, and pedogenic carbonate would be that from limestone dissolved and reprecipitated at depth. Accumulation of both these pedogenic products might be more rapid than from a granitic rock.
2. Parent-material texture exerts a major control on soil formation mainly because fine-grained materials have such a high surface area per unit volume relative to coarse-grained materials. Compared to coarser grained materials, finer grained ones will:
   (a) wet to shallower depths and therefore have shallower pedogenic accumulations,
   (b) retain water longer through the year,
   (c) weather more rapidly,
   (d) require more material (Fe, carbonate, *etc.*) to coat their grain surfaces to meet identical colors in coarser grained materials, and
   (e) require longer times to reach particular carbonate morphological stages.
3. Many studies have shown the importance of atmospheric dust on pedogenesis. For example, Muhs (1983) demonstrated the influence of dust from the Mojave Desert of southern California on the soils of San Clemente, an offshore island, and the Sahara as a source for Caribbean soils (Muhs and others, 1990). Yaalon's (1987) transect across an area of varying dust input from a desert source is perhaps one of the more striking examples. For desert loesses, the main source areas are dry river beds, playas, or ancient lake bottoms, and commonly the influx of dust from these sources can be episodic due to climatic change cycles (**M** *in* Fig. 3-4). Adjacent to the source area, dust can be assimilated into the soils (**N** *in* Fig. 3-4). In other places, however, perhaps where the accretion rate is higher (**O** *in* Fig. 3-4), the soil is partly cumulic (soil formation continues as material accumulates at the surface), and some soil features resulting from past, differing soil-moisture regimes can overlap (McFadden and others, 1986). At still higher accretion rates, pedogenesis cannot keep pace during intervals of rapid but episodic deposition, as evidenced by buried soils between loess sheets (**P** *in* Fig. 3-4). A soil-loess sequence in southeastern Idaho (Pierce and others, 1982) is an example of the latter. Finally, at great distances from the source area, clay-rich atmospheric dust is assimilated into soil profiles (**Q** *in* Fig. 3-4). Because the latter area is relatively humid, pedologic evidence for episodic differences in accretion rate may be masked by pedogenesis.

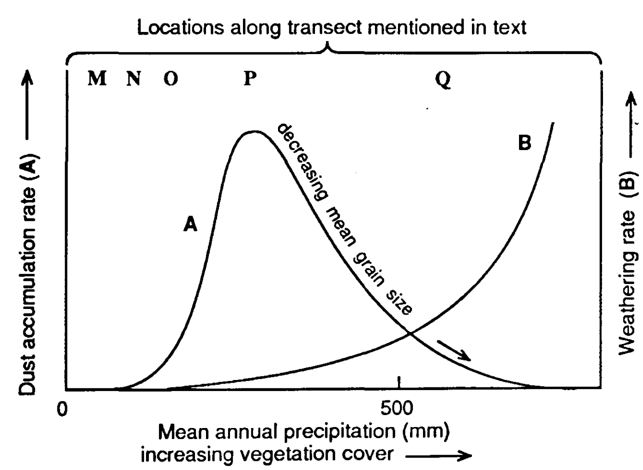


Figure 3-4. Schematic diagram showing relations between dust accretion rate (A) and weathering rate (B) along a climatic transect from a desert dust-source area downwind to areas of progressively greater precipitation and vegetation cover (modified from Pye, 1987).

CHAPTER IV
CLIMO-CHRONOSEQUENCES—THE PROGRESSIVE DEVELOPMENT OF SOILS FROM HYPERARID TO SUBHUMID CLIMATES

INTRODUCTION

Soil chronosequences are the sequential variation in soils or particular soil properties on landforms and deposits of different age. Here we will investigate such changes along a climatic gradient, from hyperarid to subhumid; hence, these might be better termed climo-chronosequences. Following Jenny (1980), quantification of the property *versus* age relations is a chronofunction.

Vreeken (1975) categorized chronosequences into four different types. The most common type is the "post-incisive sequence," an example of which is a stepped river-terrace sequence where each soil in the sequence begins to form when the terrace forms. It is assumed that younger soils progress to the older soils in the sequence, so a plot of a soil property *versus* age tracks soil development (Fig. 4-1). Examples are given in Birkeland (1984).

The interval over which a chronosequence can be traced back in time is related to how long the landscape can survive. For example, consider a series of alluvial fans in a desert basin (Fig. 4-2). The younger soils are formed on relatively flat surfaces, and with time the surfaces undergo change from their initial relatively rough bar-and-swale topography to a smooth desert pavement. Still older surfaces might be undergoing slight dissection, sufficient to erode the upper parts of the soils. Finally, the ballena stage can be reached in which the landscape is rolling, all original depositional surfaces are gone, and the soils are highly eroded. Obviously, in this case soil development will progress until erosion is more rapid than soil development, then the apparent development reverses as the soil thins (Johnson and Watson-Stegner, 1987). The time required to attain the ballena landform varies from place to place.

Bockheim (1980) selected properties of chronosequences for 27 areas from the literature, and correlated properties with time using regression techniques and three models:

$$Y = a + bX,$$
$$Y = a + (b \log X), \text{ and}$$
$$\log Y = a + (b \log X),$$

where Y is the soil property and X is time. Correlation coefficients were calculated, as were their statistical significance. The second model, $Y = a + (b \log X)$, yielded the highest correlation coefficients for most soil properties, and 85% of these coefficients were statistically significant. This technique has been used by many subsequent workers. Recently, Reheis and others (1989) suggested that linear plots might be indicative of the input of eolian materials, whereas log plots could be indicative of weathering related features (*i.e.*, rubification). Finally, Switzer and others (1988) discussed statistical methods useful in making such plots.

Data in this manual are portrayed several ways (Fig. 4-3). One is a plot of the content of some soil constituents *versus* depth, the most common portrayal of soil data. Another way

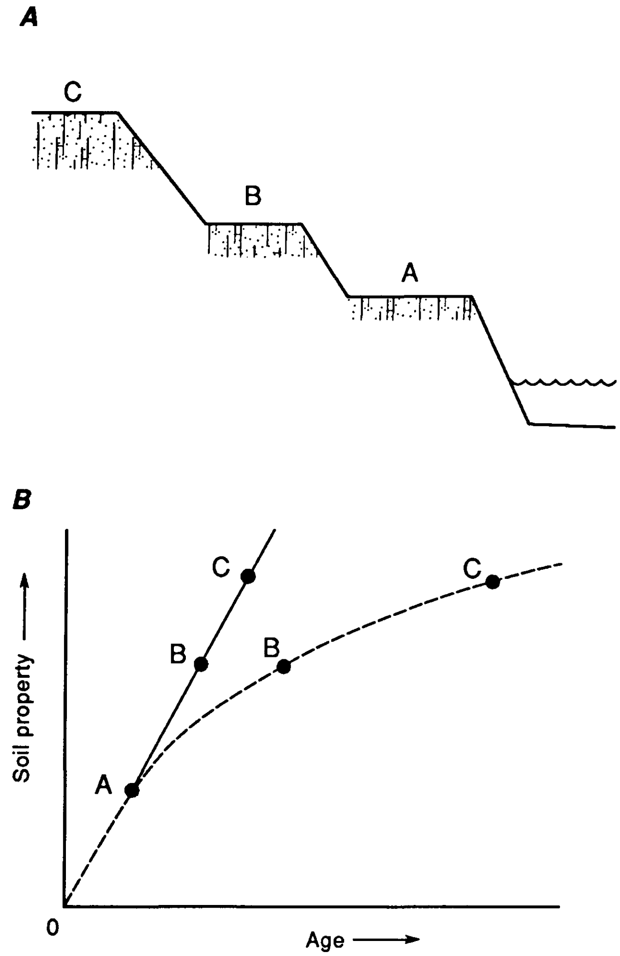


Figure 4-1. ***A.*** Example of a post-incisive soil chronosequence (soils A, B, and C) formed from river-terrace deposits of different ages. ***B.*** Data (*e.g.* profile-development index, g of clay/cm2/profile, or other properties) for soils A, B, and C, plotted against age of the river-terrace deposits. A linear function (solid line) might define a minimum age relation or a property controlled by constant eolian influx, whereas a logarithmic or power function (dashed line) seems best for many soil properties *versus* age and might define a maximum age relation (see Colman and Pierce, 1981).

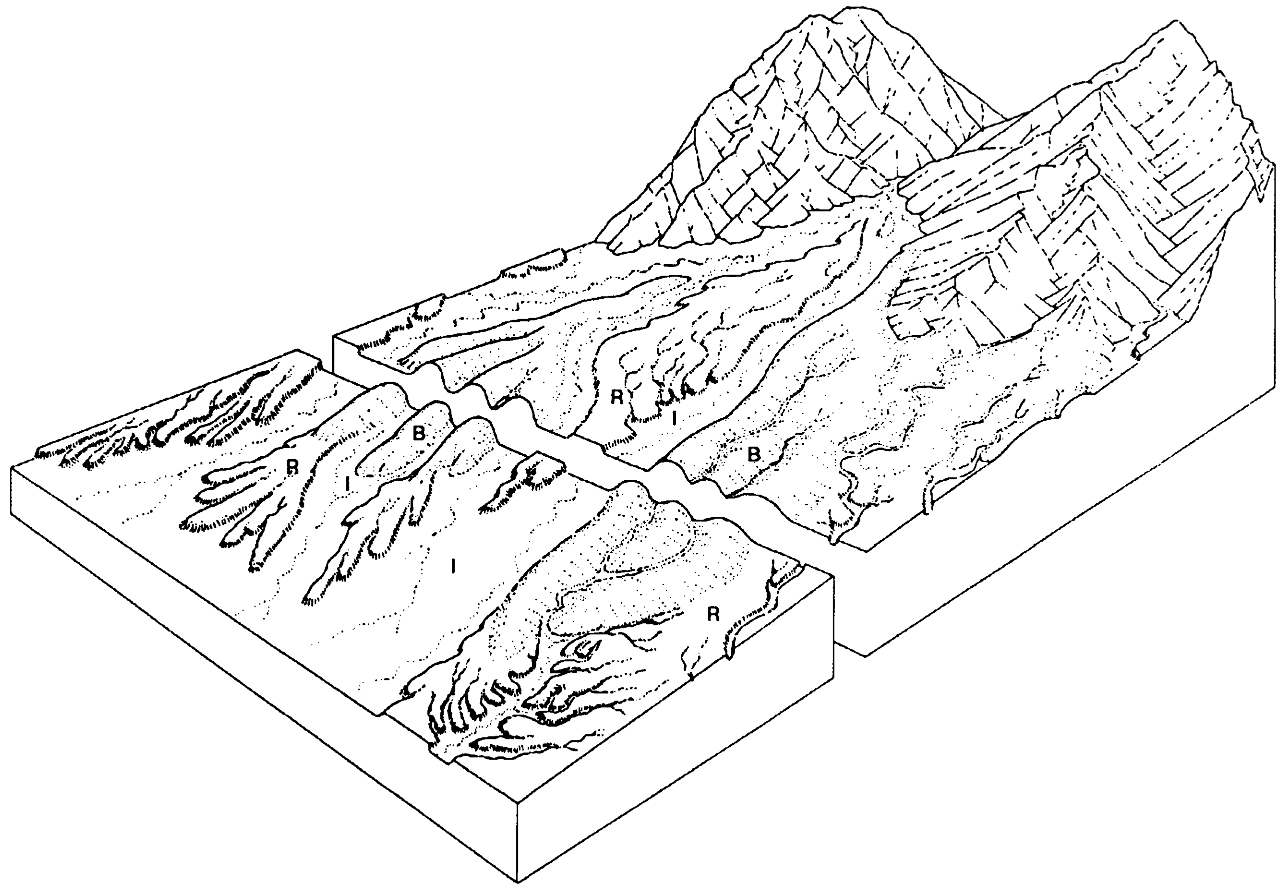


Figure 4-2. Schematic diagram of alluvial fans at the front of a desert mountain range (from Peterson, 1981). The youngest surfaces (I) are connected to the drainages and little dissected. The next older surfaces are remnants (R) isolated from the parent stream, and slight dissection of the surfaces is shown by the drainage pattern on the surfaces. As erosion proceeds, all original topography is lost and eroded ridges (B, ballenas) are all that remain.

is the accumulation index of Levine and Ciolkosz (1983) in which the parent-material content is subtracted from the horizon content of a particular soil constituent, the result is multiplied by the horizon's thickness, and the horizons are summed for a profile value. In Figure 4-3, the accumulation index is represented as the shaded area to the right of the parent-material value. One should correct these values for the volume of the soil occupied by gravel. A third way is the mass accumulation in which the percent of the pedogenic material accumulated (parent material corrected) for a horizon is multiplied by the bulk density, the latter is multiplied by the horizon thickness, and all horizons are summed; this gives a profile value in g/cm^2 of landscape surface area for the described profile. The gravel content will alter the true value of each of the above portrayals of soils data and should be taken into consideration in any comparative studies.

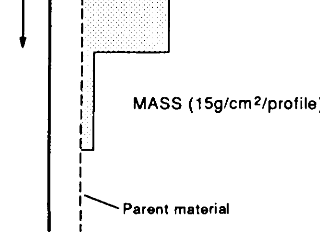


Figure 4-3. Three examples of the ways in which soil data are portrayed: a graphic plot, the accumulation index (AI), and the mass accumulation.

CHRONOSEQUENCES IN ARID AND SEMIARID REGIONS

Salts accumulate in soils in areas of limited effective soil moisture. For example, in southern Israel carbonate accumulates in soils at less than about 50 cm mean annual precipitation (MAP) and gypsum at less than about 25 cm MAP (Dan and Yaalon, 1982). Still lower MAP is required for halite accumulation. In the following discussion, MAP and MAT (mean annual temperature) values are given in centimeters (cm) and degrees centigrade (°C), respectively.

Rapid soil development takes place in the extremely arid environment of southern Israel (Dan and others, 1982), where much of the pedogenic salt, silt, and clay are considered to be of atmospheric origin (Gerson and others, 1985; Gerson and Amit, 1987). An example is soils on a flight of river terraces that range from recent to an estimated age of 14 ka (MAP 3; MAT 25) (Amit and Gerson, 1986; Gerson and Amit, 1987). After about 14,000 yr of soil formation, coarse-clay content approaches 10% and fine clay 15%, fine silt accumulation is a maximum of 25% near the surface, and gypsum is about 8% (Fig. 4-4). For older soils in the region with estimated ages of several hundred thousand years, maximum silt content is 43% near the surface and 24% at depth in the B horizon, and the gypsum maximum of 43% produces a Bym (petrogypsic) horizon (Dan and others, 1982; Gerson and Amit, 1987).

A chronosequence of alluvial-fan deposits in north-central Wyoming (MAP 7; MAT 7) suggests rates of clay and gypsum accumulation somewhat similar to those of southern Israel (Reheis, 1987a). Marked clay accumulation (27%) and some gypsum (0.2%) are noted in 5-ka soils, and gypsum exceeds 50% in some horizons of soils estimated to be 410 ka.

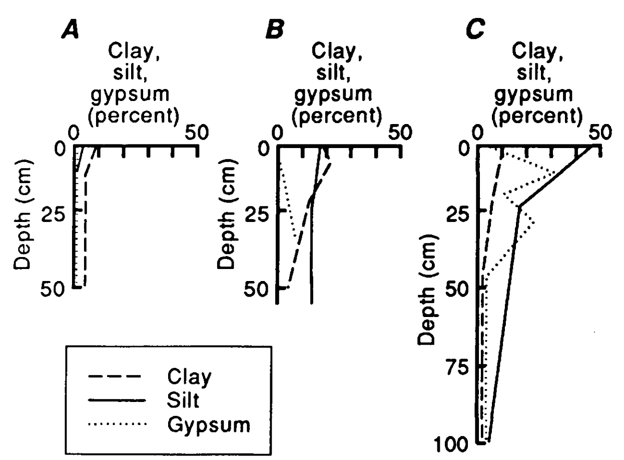


Figure 4-4. Data on clay, silt, and gypsum contents in soils of various ages in southern Israel (from Dan and others, 1982; Amit and Gerson, 1986). ***A.*** Late Holocene. ***B.*** 14 ka (latest Pleistocene). **C.** Middle Pleistocene.

Much research on soils has been undertaken in the Western United States. In general, pedogenic $CaCO_3$ accumulates in soils where MAP is less than about 50 cm, similar to the value for Israel. The key work is that of Gile and Grossman (1979) and Gile and others (1981), who made many significant contributions to arid-region soil geomorphology. First, much of the material accumulating in the soils is of atmospheric origin. Hence, 5*YR* Bt horizons can form in as little as several thousand years in gravelly materials, a rate considered too rapid if weathering processes are responsible for all of the clay (Fig. 4-5). The red color, however, could reflect Fe released by weathering. Second, the carbonate morphology stages change in such a predictable manner with time that they can be a powerful tool in the correlation of deposits. Finally, detailed maps are presented to demonstrate the relation of soils to landscape units; of importance is the fact that subtle variations in soils can be used to identify areas of erosion not always readily seen on surface examination alone. These maps are ideal models for understanding complex soil-geomorphic relations.

Machette (1985b) added to the work of Gile and others (1981) in a regional study that demonstrated that the time for attainment of a particular carbonate stage varies between

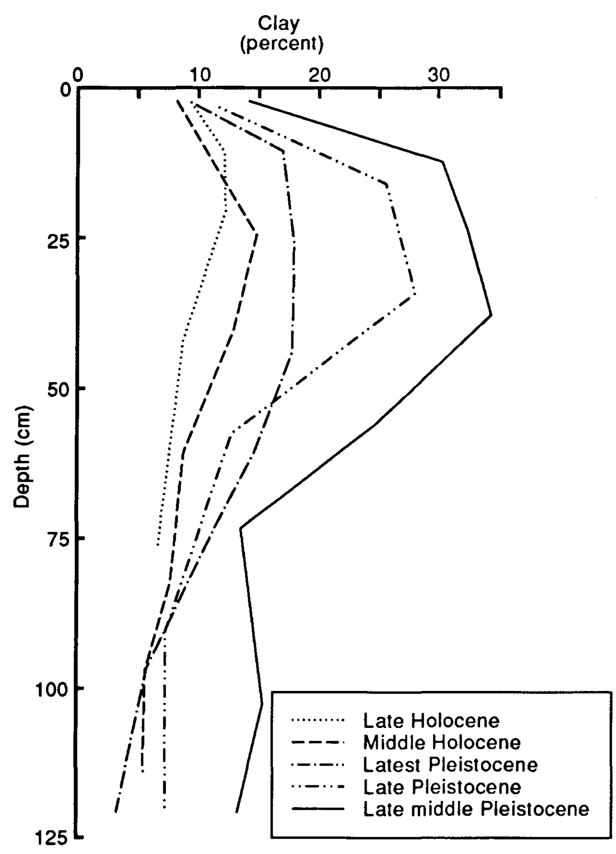


Figure 4-5. Clay distribution in gravelly soils near Las Cruces, New Mexico (from Gile and Grossman, 1979).

areas; low rates are found in either mountains or basins where atmospheric dust influx is low and the dust has a low carbonate content, and high rates under the opposite conditions (Fig. 4-6). By examination of stable soils of known age (parent materials dated by associated volcanic ash), he estimated the rate of mass accumulation of carbonate, and found a positive correlation between the accumulation rate and the rapidity with which successive carbonate morphological stages are attained. This method is preferred over one that determines accumulation rate by dust traps placed in the landscape (Gile and others, 1981), because the former is a measure of what actually gets into the soil in the long term.

There are some well established carbonate influx rates for Utah. Harden and others (1985) obtained a range of rates from 0.14 to 0.26 $g/cm^2/10^4$ yr for the Spanish Valley, north of the La Sal Mountains, and Scott and others (1983) estimated a rate of 0.5 $g/cm^2/10^4$ yr for the eastern Lake Bonneville area. Machette (1985b) estimated a rate of 0.14 $g/cm^2/10^4$ yr for the Beaver basin, although some soils (pedalfers) at higher altitudes (>1,950 m) have been leached of carbonate.

A fairly well-dated chronosequence consisting of both surface and buried soils exists at the Lubbock Lake Archaeological Site, Texas (MAP 47; MAT 15) (Holliday, 1988). The youngest soil formed at the surface in the past 100 yr and where best expressed has an A/Bwk profile with stage I carbonate morphology. The intermediate-age soil formed over 450 yr, where it is at the surface, and 200 yr where buried; the best expressed profile is a A/Bt/Bk with stage I carbonate morphology. In contrast, the oldest soil had 4,500 yr to form where not buried and 3,500 yr to form where buried, and the best expressed profile is A/Bt/Bk with a maximum carbonate morphology of stage II. Parent-material hue is 10*YR*, and the oldest soils have 7.5*YR* to 5*YR* hues. An adequate amount of dust and moisture combined to produce the exceptionally rapid development of the Bt horizon, and a rate of carbonate accumulation that could be as much as 10 times faster than the rapid rates for soil development in areas of eastern New Mexico (Fig. 4-6).

Soils formed on alluvial deposits derived from rhyolitic ash-flow tuffs at the Nevada Test Site (MAP 15; MAT 15) accumulate secondary silica in addition to carbonate (Taylor, 1986). Pedogenic opal-silica stages are defined, modified after the carbonate morphology stages. Atmospheric dust seems to account for the accumulation of clay and carbonate, whereas weathering seems to have released sufficient silica to form the pedogenic opal silica. Soils of about 10 ka age lack a distinct Bt horizon but have a 10*YR* Bw horizon, stage I-II carbonate morphology and stage I silica morphology. The youngest deposit with a Bt horizon is 30-47 ka. By 270,000-430,000 yr, the Bt is 7.5*YR* hue, a K horizon is present, the carbonate and silica stages are III-IV and III,

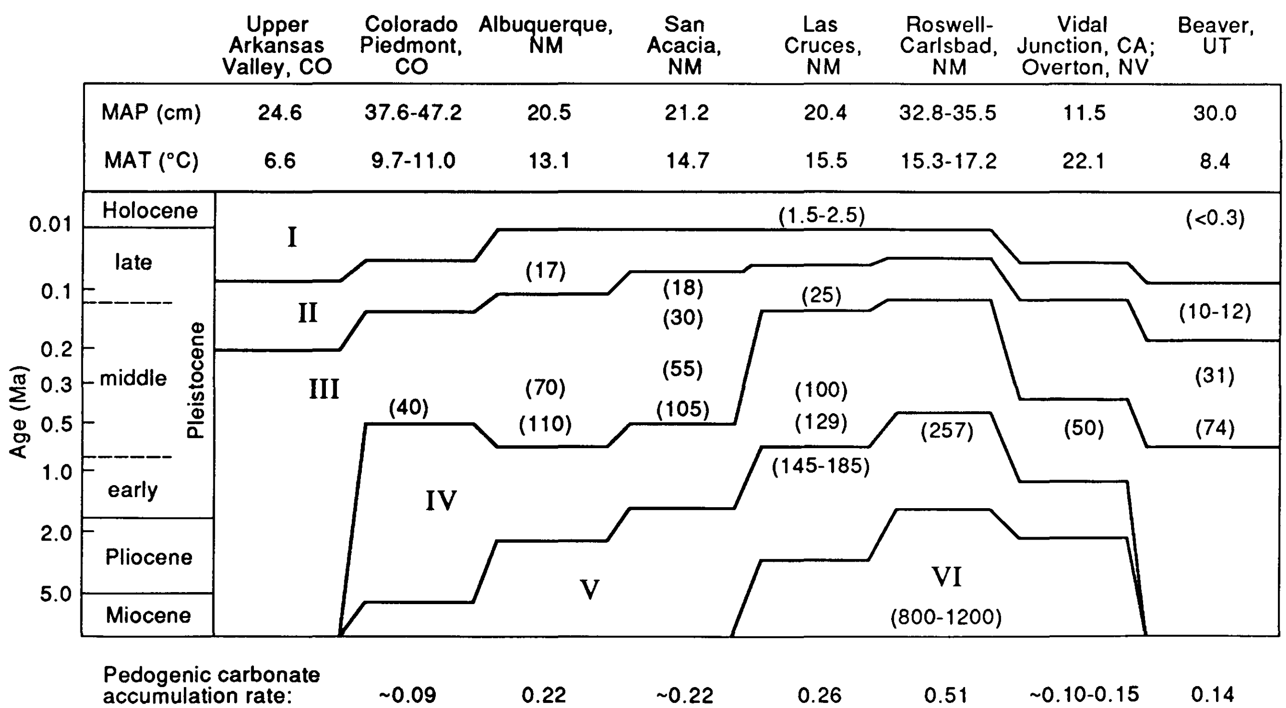


Figure 4-6. Maximum stages of carbonate morphology in relict soils formed in gravelly alluvium in the southwestern United States. Numbers in parentheses are average weight (g) of $CaCO_3$ in a one cm^2 column of soil. Pedogenic carbonate accumulation rates are $g/cm^2/10^3$ yr for the past 500,000 yr (data from Machette, 1982; Machette, 1985a; and McFadden, 1982).

respectively, and there is a distinct relation between the development of various properties and elevation (Fig. 4-7). The long-term rate of carbonate accumulation is 0.1-0.3 $g/cm^2/10^3$ yr, a moderate rate compared to those elsewhere in the southwestern United States (Fig. 4-6), and that of pedogenic silica is 0.08 $g/cm^2/10^3$ yr.

The work of Gerson and Amit (1987), Reheis (1987a), Wells and others (1987), and Chadwick and Davis (1990) can be combined to suggest the following long-term soil-geomorphic response of gravelly fluvial deposits in arid lands. Shortly after a flood plain is abandoned, bar-and-swale topography is present, and such a rough surface is an ideal dust trap. In time, weathering reduces the size of surface rocks, colluviation and eolian influx erode the bars and bury the swales, and the bar-and-swale topography is replaced by a smooth surface with desert pavement. A recently evoked origin for the desert pavement is that atmospheric silt and clay infiltrate beneath the surface stones to form a gravel-free Bt horizon and at the same time lift the stones so that the latter always remain at the surface (McFadden and others, 1987). The smooth surface is a less efficient dust trap than the rough one, so that with time a greater proportion of the atmospheric solids does not get translocated to depth but bypasses the site. The formation of a surface crust, primarily in the Av horizon, and the plugging of voids with fines diminishes infiltration and the capacity of the soil to transmit solids vertically in suspension and salts in solution; high concentrations of Ca^{2+} could flocculate clay particles and limit the depth to which they are carried. One result of decreasing soil-infiltration capacity would be relatively greater runoff, the development of a drainage network on the abandoned terrace, and eventual erosion of the soil.

In some extremely arid environments with both gypsum and carbonate available in dust, pedogenic gypsum accumulation seems to dominate (Reheis, 1987a). One reason for this is that gypsum is the more soluble of the two substances; therefore, more of it might be dissolved and carried into the soil. In contrast, the undissolved carbonate at the surface could be removed by eolian transport. A second reason is the common ion effect, with the high concentrations of Ca^{2+} from gypsum dissolution markedly inhibits $CaCO_3$ solubility.

A final word of caution on the rates of various processes in arid environments. The position of a soil profile on extensive fan surfaces is important because soils closer to dust sources will develop more rapidly than those at farther distances from the same sources (Reheis and others, 1989; Chadwick and Davis, 1990).

CHRONOSEQUENCES IN SUBHUMID REGIONS

At effective soil moistures higher than those at which $CaCO_3$ accumulates, soil chronosequences have a morphology dominated by the accumulation of clay and the accumulation or depletion of various chemical constituents. Here, we will discuss mainly chronosequences in the mountains and high-altitude basins of the Western United States; the

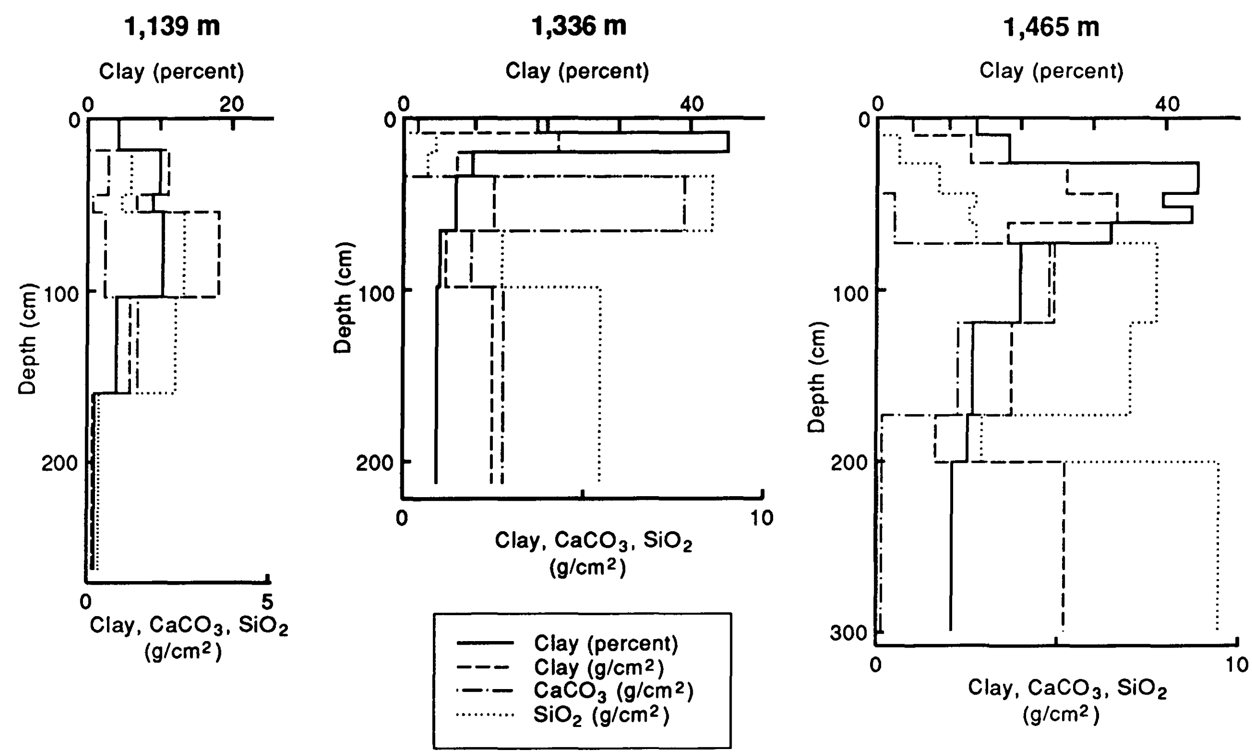


Figure 4-7. Clay, carbonate, and silica contents for soils about 270-430 ka formed from river-terrace deposits at different elevations at the Nevada Test Site (modified from Taylor, 1986).

mountains are relatively moist, and effective soil moisture decreases with decreasing altitude. Richmond's (1962) work in the La Sal Mountains, Utah, still stands as a model for such work. Part of his work has been revised and quantified by Harden and others (1985). Climate and availability of $CaCO_3$ in dust probably dictate whether or not the soils are calcic in these areas.

Holocene chronosequences have been reported for many high-altitude cirque moraines. In general for the Western United States, soils have well-differentiated A/Bw/Cox or A/Btj/Cox profiles in the areas of higher rainfall after 10,000 yr of formation (Mahaney, 1978; Shroba and Birkeland, 1983; Birkeland and others, 1987). In places, the accumulation of clay and Fe(d) are more rapid than in soils formed from 20-ka moraines adjacent to the mountains at lower elevations (Shroba and Birkeland, 1983; Birkeland and others, 1989).

Soils formed in till on moraine crests of major glaciations have been studied at numerous mountain fronts in the Western United States, from the Sierra Nevada on the west (MAP 40-45; MAT 7-8) to the Rocky Mountains on the east (MAP 40-90; MAT -1 to 6). Clay and Fe accumulation, as well as rubification (reddening), increase at low rates across the region. Perhaps the lack of atmospheric dust and low effective soil moisture at most localities account for these low rates. In general, soils on 20-ka tills have A/Cox and A/Bw/Cox profiles, and some have Bt horizons (Shroba and Birkeland, 1983). Colors commonly are no redder than 10*YR*, but some are 7.5*YR*, clay contents are commonly between 3% and 10%, and in some places less, and Fe(d) can be as high as 0.5% at the surface. Soils formed on the crests of moraines about 140 ka have A/Bt/Cox profiles, where best expressed. B-horizon hues can be as red as 7.5*YR* and some are 5*YR*. Clay in the Bt horizons ranges from 9% to 20%. Fe(d) displays no trend with time in the Sierra Nevada; in places in the Rocky Mountains Fe(d) is 0.6%, about twice as abundant as it is in the 20-ka soils (Swanson, 1985).

In contrast to the above, soil development in the centers of some adjacent aridic basins is more rapid than in the moraines along the flanks of basins. One dramatic example is in the Lahontan basin, where rapid rates of silt and clay accumulation correlate with proximity to dry lake beds, which are the local sources of dust (Chadwick and Davis, 1990).

Chronosequences on river-terrace deposits have been studied at several localities in California, and the rate of development in each seems to depend strongly on the rate of dust influx. For example, the Great Valley of California (MAP 30; MAT 16) is an area of relatively low dust influx. The youngest soil with a Bt horizon is about 40 ka, the Bt is 68 cm thick, has 10*YR* hue, and clay and Fe(d) maxima of 17% and 0.5%, respectively (Harden, 1987). After about 3 m.y. of soil development, however, the Bt horizon is over 7.5 m thick, has 10*R* hue, 63% clay and 3.4% Fe(d).

In contrast, soils in areas of relatively high dust influx form at more rapid rates. McFadden and colleagues (McFadden, 1982; McFadden and Hendricks, 1985; McFadden and Weldon, 1987; McFadden, 1988) have studied soil chronosequences in a transect across southern California. Noncalcic soils in the wetter part of the transect (MAP 40-78; MAT 16) form Bt horizons more rapidly than those in the Great Valley. After about 10,000-15,000 yr, Bt horizons are about 100 cm thick, 7.5*YR* hue, 10% clay and 1.4% Fe(d). Bt horizons in soils between 300 and 700 ka can be as much as 5 m thick, 2.5*YR* or 10*R* hue, 39% clay and 2.9% Fe(d). The rate of soil development is still more rapid, however, in alluvial deposits on the California coast near Ventura (MAP 36; MAT 15) (Harden and others, 1986) where Na^+ is abundant. Bt horizons are present after 40,000 yr, they are at least 43 cm thick, have a hue of 7.5*YR*, and a maximum of 34% clay. The oldest soils in the sequence are between 80 and 105 ka and have a Bt horizon over 2.1 m thick, a 5*YR* hue, and a maximum of 28% clay.

COMPARATIVE DATA ON CLAY AND IRON BUILDUP AND PDI IN SOILS OF DIFFERENT CLIMATES

Data are available to compare clay and iron buildup in soils in various environments. However, for much of the data used here, the parent-material amounts have not been subtracted from the totals in the soils so that pedogenic amounts cannot be reported. This distinction is important because pedogenic amounts should plot at the origin for 0 yr. In contrast, if parent-material amounts are not subtracted from the data, it is difficult to know the shape of many curves in the first 10,000 to 20,000 yr of soil formation.

The bulk of the clay-accumulation data seems to fall into two artificial groups (Fig. 4-8). High amounts (group A) are characteristic of the coast and Great Valley of California. In contrast, low amounts (group B) are characteristic of the Sierra Nevada, the arid and semiarid soils of McFadden's (1982) southern California transect, Rock Creek, Montana, and the Rocky Mountains. The xeric soils of McFadden's (1982) transect lie between the two groups.

The Fe(d)-accumulation data also fall into two artificial groups (Fig. 4-9). Greater accumulation (group A) is present in Ventura, California, and the xeric soils of McFadden's (1988) transect. Lower accumulations (group B) are present in the Great Valley, the semiarid and arid soils of McFadden's (1988) transect, and the Rocky Mountains (Swanson, 1984, 1985).

Factors involved in the above trends in clay and Fe(d) are the parent-material amount, atmospheric influx, and weathering of various primary minerals. The relative importance of these factors varies with the area, and is not always addressed by workers. Two additional general comments can be made about these data. One is that, in general, the soil chronosequences with high clay accumulation also have high Fe(d) accumulation; this supports the findings of McFadden (1988), only here the data are for a wider range of environments. Secondly, one would intuitively expect that wetter environments (see Birkeland, 1990) have much higher amounts of Fe(d) compared to that in the relatively dry soils of the Western United States, but this does not appear to be the case.

PDI values increase in most areas with time. Intuitively, one would think that plots of PDI *versus* age would be different in different climatic regimes. However, in one study that included climates that range from arid to humid most of the data cluster broadly together (Fig. 2-6; Harden and Taylor, 1983). The clustering was best when only the four best field properties in each area were used. This has important implications when field soil data are used to provide broad age estimates for deposits. However, one might also plot all properties *versus* age, as it is important to know if PDI *versus* age trends differ in different climatic regimes.

Finally, trends such as those shown here demonstrate that soils can be used to estimate the ages of deposits. Unfortunately, ± 50% is probably the best error on soil-derived age estimates that one can hope for. In a statistical study of PDI values, Harden (1987) put the error limits at ± 70%. For particular parts of the time scale such estimates are quite valuable, and perhaps no worse than age estimates by some other relative-age techniques (*i.e.*, incision rate, rock varnish cation ratios, ^{14}C dating of rock varnish, weathering-rind data, *etc.*).

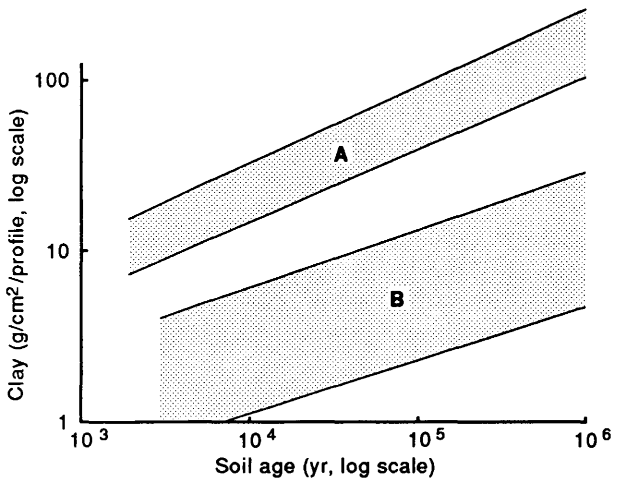


Figure 4-8. Clay accumulation with time in various environments (from Birkeland, 1990). Group **A** includes soils from San Clemente Island, California (Muhs, 1982a); Ventura, California (Harden and others, 1986); and the San Joaquin Valley, California (Harden, 1987). Group **B** includes soils from the Rocky Mountains (Swanson, 1984; Colman and Pierce, 1986); Rock Creek, Montana (Reheis, 1987b); arid and semiarid areas of southern California (McFadden, 1982); the Sierra Nevada, California (Berry, 1990); 20-ka soil in the San Luis Valley, Colorado (McCalpin, 1982). Xeric soils (not shown) of southern California (McFadden, 1982) and 120- and 400-ka soils in San Luis Valley, Colorado (McCalpin, 1982) plot at intermediate values. Only data for Ventura, San Luis Valley, and Rocky Mountains have been corrected for amount of parent-material clay. Data for humid-area soils are included in the plots; see Birkeland (1990) for locations.

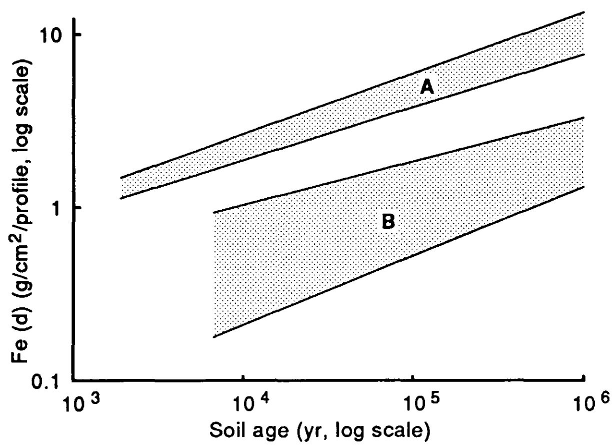


Figure 4-9. Fe(d) accumulation with time in various environments (from Birkeland, 1990). Group **A** includes soils at Ventura, California (Harden and others, 1986) and xeric soils of southern California (McFadden, 1988). Group **B** includes soils of San Joaquin Valley, California; arid and semiarid soils of southern California (McFadden, 1988); and the Rocky Mountains (Swanson, 1984, 1985; Colman and Pierce, 1986). Only data of McFadden (1988) have been corrected for amount of parent material Fe(d). Data for humid-areas soils are included in the plots; see Birkeland (1990) for locations.

CHAPTER V
TOPOGRAPHY-SOIL RELATIONS

INTRODUCTION

Topography, or local relief, controls the distribution of soils in the landscape to such an extent that soils of markedly contrasting morphologies and properties can merge laterally with one another and yet be in equilibrium under existing local conditions. Many of the differences in soils with varying topographic settings are due to combinations of microclimate, pedogenesis, and geological surficial processes, and differentiating the effects of each on soil distribution is difficult. In assessing the topographic factor, the fields of pedology and geomorphology probably overlap more than with any other pedologic factor (see discussion in Gerrard, 1981).

Examples of how soil properties vary laterally with topography are plentiful (Ollier, 1973; Jenny, 1980; Gerrard, 1981; Birkeland, 1984). In arctic climates, well-drained A/Bw profiles on slopes might grade downslope to cryoturbated, poorly drained soils. In semiarid terrain, A/Bw profiles on slopes might grade downslope to A/Bt/Bk profiles. Finally, in tropical areas, red kaolinitic soils might grade downslope to dark-colored smectitic soils.

There are many reasons for the downslope variations in soil profiles (Fig. 5-1). One reason for this is the orientation of the hillslopes on which soils form; this affects the microclimate and, hence, soil genesis. Another reason is the steepness of the slope; this affects soil properties because the rates of surface-water runoff and erosion vary with steepness and length of the slope. In areas of rolling terrain, soil properties may vary because lower areas are likely to be areas of accumulation of water runoff, throughflow (water moving laterally through the soil), and sediment derived from surrounding higher lying areas; in places, buried soils might be present. Also, low areas might be influenced by a high water table, which can result in mottled or gleyed soils.

THE CATENA

Numerous studies have shown that many soil properties are related to the gradient of the slope, as well as to the particular position of the soil on a slope. Milne (1935a, 1935b) proposed the term catena (from catenary curve) to describe the lateral variability of soils on hillslopes and emphasized that each soil along a slope is linked to and bears a distinct relation to the soils above and below it for a variety of geomorphic and pedologic reasons. These catenas are also called toposequences. Yaalon (1975) has reviewed many of the topofunctions derived from various areas in the world, and Muhs (1982b) has a more recent review focussing on catenas in Mediterranean climates.

Several models have been proposed to describe how landscapes are related to soil catenas. Conacher and Dalrymple (1977) proposed a nine-unit landsurface model defined on the basis of process and response. For many studies, however, it is sufficient to use a simple five-unit model based on slope form. Going from the top to the base of a slope (Fig. 5-1), these units are (1) the flat summit (Su), (2) the convex portion or shoulder (Sh), (3) the more or less uniform slope of the backslope (Bs), and the concave portion at the base, which is divided into (4) an upper footslope (Fs) position and (5) a lower toeslope (Ts) position. Soil catenas can be further divided into open and closed systems (Ruhe and Walker, 1968; Walker and Ruhe, 1968). In an open system, drainage is such that sediment can leave the area. The closed system is characterized by a closed depression, and all eroded sediment is trapped in the depression.

Although most soil catenas are depicted as a two-dimensional cross section, Huggett (1975) suggested that a better approach would be to consider a three-dimensional model, such as a small watershed. The watershed's boundary is the divide with the adjacent watershed. He pointed out that properties of soil catenas in such a setting could be partly explained in terms of lateral flow of soil water, called throughflow water, from the higher parts to the lower parts

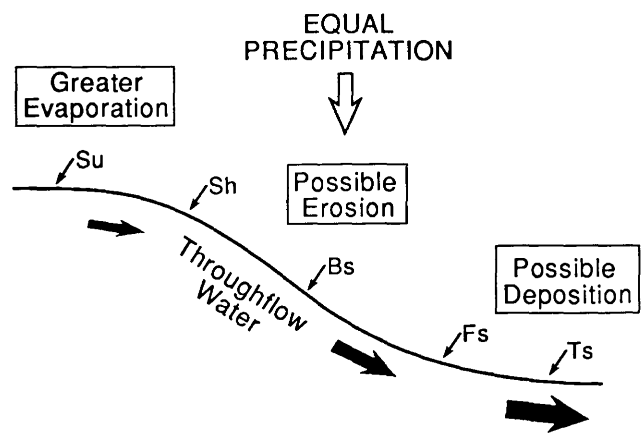


Figure 5-1. Schematic diagram of some geomorphic and pedologic processes related to position on a slope. Sites commonly studied are in the summit (Su), the convex part or shoulder (Sh), the straight section or backslope (Bs), the concave part or footslope (Fs), and the relative flat part beyond the footslope (Ts, the toeslope). Although precipitation is equal at all sites, evaporation is greatest at the summit. In more humid environments, water is redistributed downslope both as surface runoff and as throughflow water, so that the lower parts of the slope receive more water (indicated by width of arrow for throughflow water). Soil properties should correlate with erosion-deposition patterns and relative amounts of soil water.

of the watershed (Whipkey and Kirby, 1978). Referring to the three-dimensional model in a humid environment, one can identify three zones based on soil chemistry: (1) in the upper reaches, an eluvial zone in which soils may show a net loss; (2) along the lower parts of the watershed, an illuvial zone where the same constituents might accumulate; and (3) a transluvial zone separating the eluvial and illuvial zones. The degree to which the soil properties match the zones will be mainly a function of the climatic regime, pH, and redox potential. Clay particles can also be moved laterally by throughflow water (Huggett, 1976).

Several kinds of catenas are recognized (Fig. 5-2). One is formed on bedrock, but this is not a very satisfactory catena because there is no certainty that either parent material or time are constant (equivalent) at all sites. For example, there could be pockets of old soil in places along the slope. The nine-unit landsurface model might be used in these cases. A second kind of catena is one formed from surficial materials that were deposited at about the same time. An example would be a till in a moraine or sand in a dune. This type of catena has the advantage over the bedrock catena in that parent material and time can be constant at all sites. Another catena formed in surficial materials is a fluvial scarp within a flight of river terraces. The terrace (flat) and the terrace scarp (riser) above it are usually about the same age, and if the fluvial materials are nearly the same texture and lithology, then the parent material and time factors can be considered as constants. The next higher terrace, however, is older and can have an older soil related to it. Using river terraces, one can create a synthetic catena in which the soil on the lower (younger) terrace is the summit soil for the catena (Fig. 5-2). Thus, the five-unit land surface model, or some modification of it, is sufficient for describing catenas formed in surficial materials.

The selection of catena sites on a slope must be made with great care. The soils in the catena should be aligned so that the surface or throughflow waters at one site have the potential for passing through all other sites. In other words, the soil pits are aligned along the fall line of the slope (*i.e.*, at right angles to the contours). With this placement strategy, the soils are linked in a physical and geochemical sense, so that one can predict what might be expected in soil formation from one site to the next.

Soil geochemistry may vary with slope position if there is sufficient rainfall to bring about significant redistribution of soil water along the slope. Elements released by weathering may or may not be redistributed along the slope as a function of their mobility in the constant or changing geochemical system along the slope (Table 5-1). For example, under slightly acid-oxidizing conditions, Fe and Al are relatively insoluble and will accumulate close to the point of release, whereas the more mobile cations (Ca, Mg, Na, and K) can move to the lower parts of the slope. In contrast, under acid-chelating conditions, both Fe and Al can be mobilized and moved to lower parts of the landscape, as can the common cations. If conditions in the lower parts of the landscape change to one of a relatively higher pH and reducing conditions, it is possible that Fe will remain mobile and leave the system, but Al will accumulate. Finally, under relatively high pH conditions and low rainfall, all pedological constituents (including the soluble salts such as carbonate) might accumulate where introduced into the soils and not be redistributed downslope.

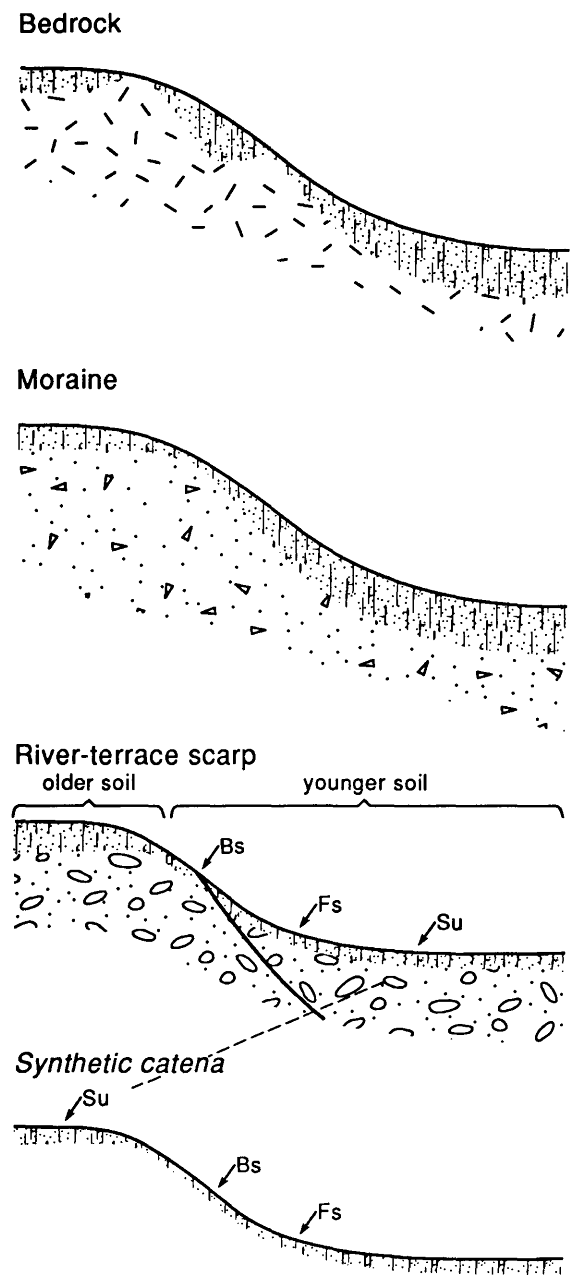


Figure 5-2. Soil development on several kinds of catenas. Depth and degree of soil development shown by line and stipple patterns. The synthetic catena is derived from the river-terrace-scarp catena.

Table 5-1.—Relative mobility of various elements with different soil conditions
[From Young and others (1977)]

Relative mobility	Mobility phase	SOIL CONDITIONS: Slightly-acid oxidizing	Acid-chelating	Reducing
↑	I	Cl^-, SO_4	Cl^-, $SO4^=$	Cl^-, $SO4^=$
	II	Ca^{2+}, Na^+	Ca^{2+}, Mg^{2+}, Na^+	Ca^{2+}, Mg^{2+}, Na^+
Increasing		Mg^{2+}, K^+	K^+, Fe^{3+}, Al^{3+}	Fe^{3+}, Mn^{2+}
mobility	III	SiO_2, P	SiO_2, P	SiO_2, P
	IV	Fe_2O_3, Al_2O_3		Al_2O_3

Clay minerals and other pedogenic products may follow the above geochemical trends if they are not in equilibrium with the soil waters and (or) sufficient time has elapsed for authigenic clay minerals to form. In general, if there is a major difference in clay mineralogy with slope position, the upslope clay-mineral suite will be more depleted in silica relative to those in the downslope position where silica accumulation typically takes place (Fig. 5-3). Tardy and others (1973) stated that, as a general rule, whatever mineral is present downslope under one climatic regime will be found upslope in the next drier climatic regime. Soluble salts will follow the same general rule and, in addition, their accumulation pattern follows their relative solubilities. Under one climatic regime, carbonate may occur only in the lower parts of the catena. However, continuing in a transect of increasing aridity, carbonate may occur throughout the catena. Under progressively still drier conditions, gypsum may first accompany carbonate in the lower catena positions only; finally, at the most arid sites gypsum and halite may occur across the catena.

Catenas in Alpine Areas

Alpine soil catenas can show marked contrasts in profile horizonation with slope position. Those in extremely cold and windswept environments (MAP 102, MAT -4; Barry, 1973) above treeline in the Colorado Rocky Mountains display morphological variation with position on rolling upland terrains (Burns and Tonkin, 1982). Topography controls soil moisture by limiting much of the snowfall to the lee sides of hills; these same sites also trap windblown silt and clay (loess). Soil distribution follows trends in snow-cover duration and in thickness of loess (Fig. 5-4). Summit and shoulder sites have little snow cover, and the resulting profiles have thin A horizons over thin Bw horizons. There is no loess at the most windblown sites, and patches of loess are as much as 8 cm thick at protected sites. Soils lower on the slopes show a relation to duration of snow cover. In late winter, snow covers much of the area, but with spring and summer melting the soils become exposed; those higher on the slopes first and those lower on the slopes last. Below the shoulder are the minimal snow-cover sites, which are char-

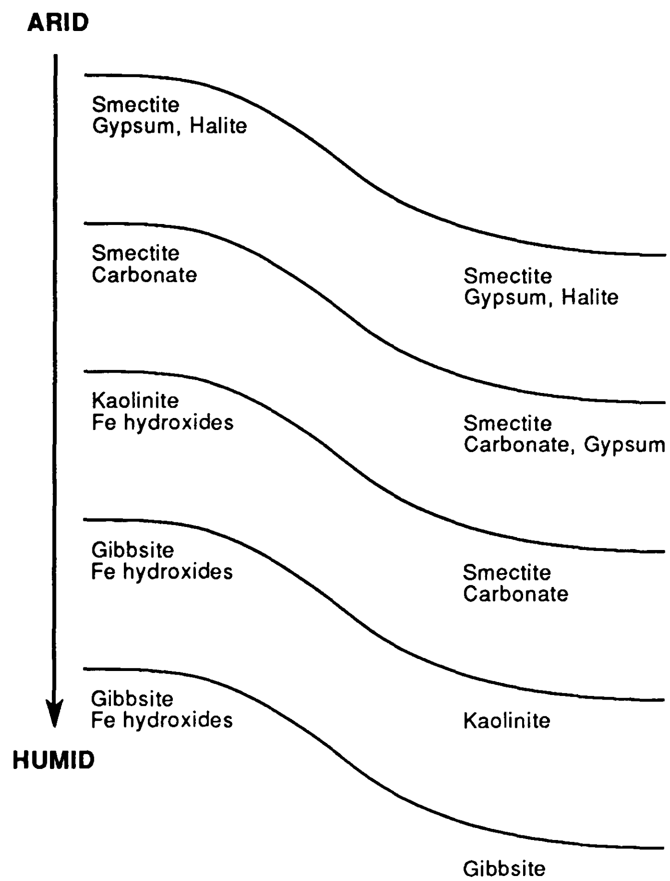


Figure 5-3. Schematic diagram showing the variation of clay minerals and other pedogenic products with position in catenas along a climatic gradient (from Tardy and others, 1973).

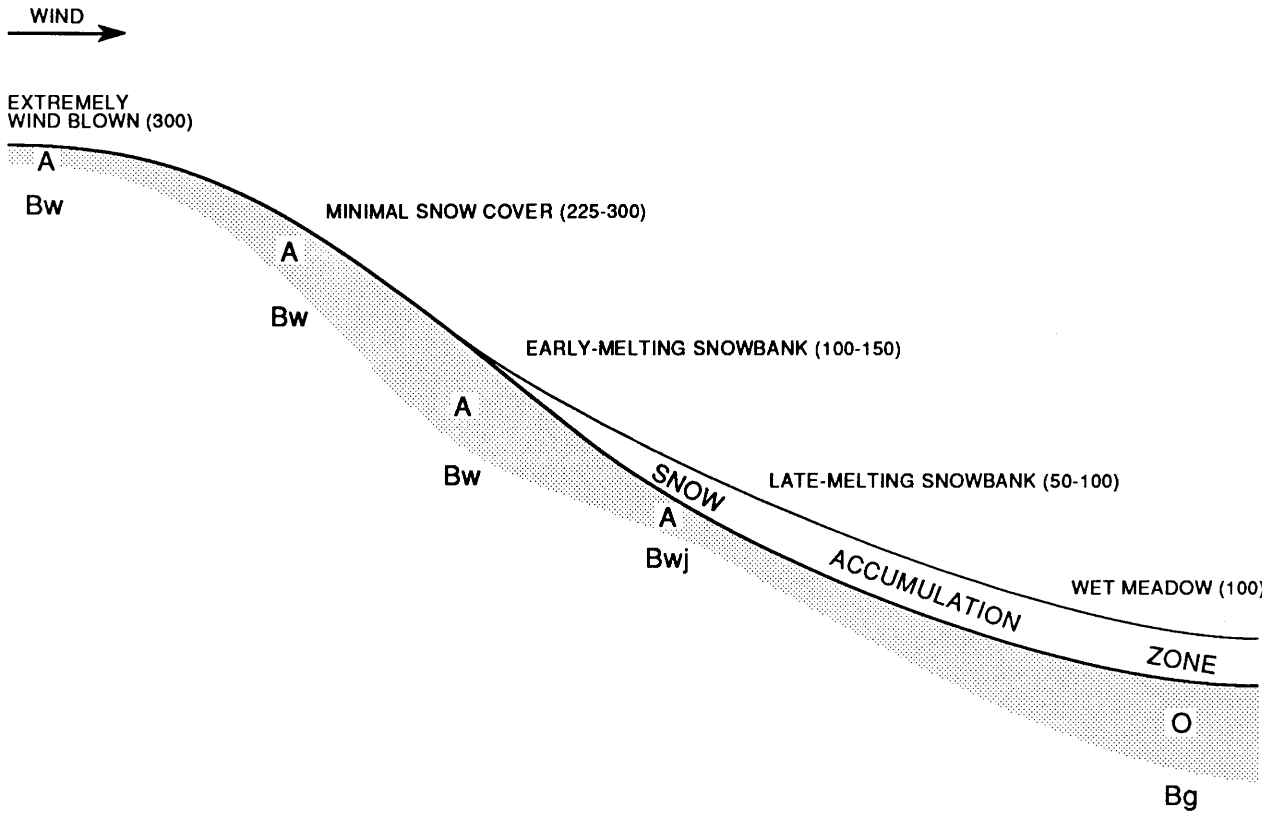


Figure 5-4. Schematic diagram of soil catena relations in an alpine tundra environment (MAP 102, MAT -4; Barry, 1973) of the Colorado Front Range (from Burns and Tonkin, 1982). Depths of A and (or) O horizon are schematic; numbers in parentheses are the estimated numbers of snow-free days per year.

acterized by the best-developed soils—thick organic matter-rich A horizons formed from loess over thick Bw horizons with the best color development in the catena. Next downslope are soils where snowbanks melt early. Loess is not as thick as in the minimal snow-cover soils, but the A horizons are the thickest of all (probably due to pocket-gopher activity) and overlie a Bw horizon. In areas where the snowbanks melt late, soils have poorly developed A/Bwj profiles with no loess. Finally, at the base of the slope are soils with poorly drained, mottled, and gleyed O/Bg profiles, which are characterized by high concentrations of silt and clay that have been transported during snowmelt from upslope positions.

Mahaney and Sanmugadas (1983, 1989) incorporated time in their studies of early Holocene alpine catenas on moraines in the Wind River Mountains, Wyoming. This area may have a climate similar to that in the Front Range of Colorado, described above. From upslope to downslope on early Holocene till, the main trends of the catena are: (1) greatest organic carbon upslope and because the lower slopes do not have cumulic profiles, there does not seem to be much downslope movement of surficial materials; (2) slightly lower pH's downslope; and (3) clay and Fe(d) content and depth of most intensive colors (at least 10*YR* 4/4) are greatest downslope, perhaps due to greater moisture as indicated by mottling.

Catenas in Dry Climates

Catenas in drier environments differ from those in wetter climates in that salts can accumulate in the drier climate soils. In these climates, the amount of soil moisture available for lateral transfer is small enough so that only the more soluble elements move laterally along the slope.

Nettleton and others (1968) described three soils formed from tonalite at various positions on a slope in southern California (Fig. 5-5*B*). They assumed that the three soils formed on slopes of about the same geomorphic age, and therefore the differences in the soils could be attributed to topographic position. Because of the warm climate (MAP 38; MAT 13) and precipitation mainly in the winter and early spring, the soils have A horizons with low organic-matter content. However, the properties of the B horizon vary markedly downslope. The soil at Vista (Fig. 5-5*C*) has only a Bw horizon, whereas the Fallbrook has a Bt horizon, and the Bonsall has an Na-enriched Bt horizon. Soil-moisture measurements were taken by Nettleton and others (1968) at various times of the year, and from these data and the 1965-1966 precipitation records, the moisture in the soils was estimated on a monthly basis (Fig. 5-5*A*). These data show that the downslope soils (Fallbrook and Bonsall) receive more soil moisture and retain their moisture longer than the upslope soil (Vista).

A marked feature of these soils is that the clay content increases downslope (Fig. 5-5*C*). Weathering of the primary minerals is greater downslope; consequently, most of the clay present can be attributed to weathering of the underlying rock at that site. The differences in weathering and clay-formation downslope are most likely due to soil moisture, which is controlled by slope position. Soils in lower slope positions can receive more moisture than those in upslope positions because of lateral movement, either at the surface or within the soil. Furthermore, clay formation enhances a soil's water-holding capacity, which in turn further accelerates clay formation.

Sand dunes along the Mediterranean coast in Israel (MAP 52; MAT 20) also demonstrate soil variations with landscape position (Dan and others, 1968). The soils on the tops and sides of the sand dunes have well-developed Bt horizons and color hues of 2.5*YR*, overlying a C horizon of low clay content. In the depressions, however, the soil at 3-m depth is nearly 60% clay and meets the color criteria for a gleyed horizon. Other notable differences below the surface from ridgetop to depression are a decrease in content of free iron oxide, a change in clay mineralogy from predominantly kaolinite to montmorillonite, and an increase in pH.

Variation in soil development on the Israel sand dunes is explained by eolian influx combined with slope processes and pedogenesis. Much of the clay in these soils is derived from eolian influx, some of which, after reaching the surface, is translocated downward in the soil. During pedogenesis, the clay minerals are thought to alter from montmorillonite to kaolinite under favorable chemical and leaching conditions. The soil in the depressions is not considered to have formed *in situ*, mainly because there are few mineral grains in the dune sand that can alter to clay. It is also suggested that some clay transfer could take place within the soil by laterally moving soil water at the top of the strongly developed Bt horizon (about 20 cm deep). Transfer of dune sediment from the slopes to the depression is not considered important because so little sand is found in the depression. Leaching conditions in the depressions are slight and thereby favor retention of montmorillonite as the main clay mineral.

Soil catenas on different age moraines that flank the mountains of the Western United States have been studied to track catena development through time. For the eastern Sierra Nevada catenas (MAP 40-45; MAT 7-8; Birkeland and Burke, 1988), a sandy colluvial wedge typically is present at the footslope sites, A/Bw/Cox profiles have formed at most sites on the 20-ka moraines, and there is little change in clay and Fe(d) downslope. In contrast, soils on moraines that could be 140 ka may not be any better developed at the summit sites than the 20-ka summit soils, but some soils downslope have Bt horizons with the highest amounts of both clay and Fe(d) for the catena. In Wyoming (MAP 40; MAT -1), Swanson (1985) found a rather similar story. For 20-ka moraines, the main morphology for all soils is A/Bw, both horizons thicken downslope, and (in the same direction) pH increases slightly, as does Fe(d). In contrast, the soil catena on the approximately 140-ka moraine has a Bt horizon at most sites and increases in thickness, and percent clay and Fe(d) downslope. Berry (1987) reported this same general pattern in the Salmon River Mountains, Idaho (MAP 90; MAT 1).

The above catena studies were conducted to determine (1) trends in catenas with time (see above), and (2) the

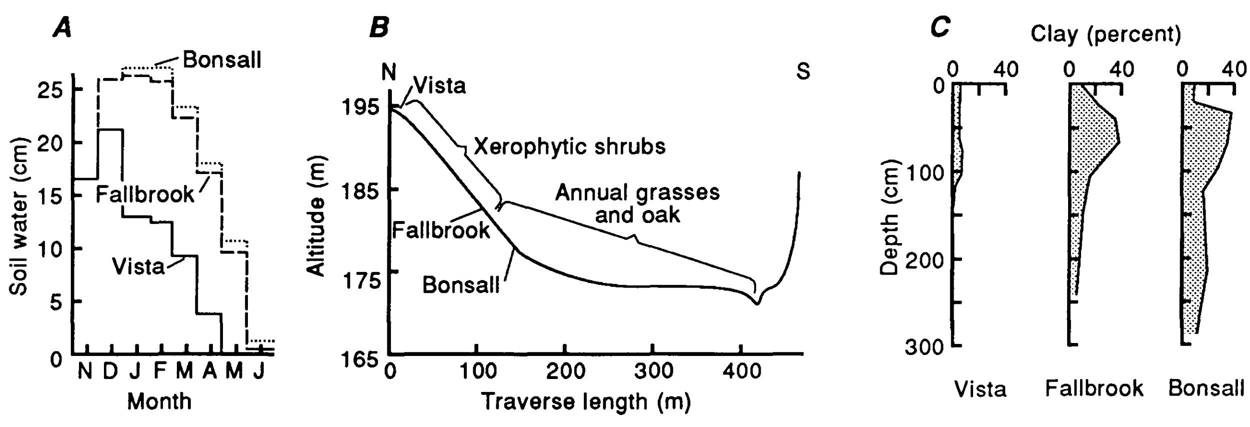


Figure 5-5. Data for a toposequence of soils formed from tonalite, southern California (from Nettleton and others, 1968). ***A.*** Estimated soil water for November 1965 through June 1966. ***B.*** Topographic and vegetative relations at the sampled sites, the slopes are 8% for Vista, 12% for Fallbrook, and 8% for Bonsall. ***C.*** Distribution of clay with depth.

usefulness of the catena concept as a field tool for differentiating moraines. For example, Burke and Birkeland (1979) clearly illustrated the problems in using summit soils solely for recognizing moraines of different ages. In some valleys, summit soils on both 20- and 140-ka moraines have nearly identical A/Bw profiles, so one must resort to other relative-age methods (*e.g.*, rock-weathering features) to differentiate the moraines in the field. If similar weakly developed soils are present at summit sites of moraines of different ages (owing to erosion), the next soil-sampling strategy is to either study multiple soils along the crests, or study soils downslope in a catena. Burke and Birkeland (oral commun., 1990) now believe the second strategy is the best one, because strongly developed Bt horizons commonly are present at some or most downslope sites in the older catenas.

PDI's can be used in the field assessment of soil catenas for making broad correlations of moraine sequences. PDI data for the eastern Sierra Nevada study indicate that the highest values generally are for the soils in downslope positions (Birkeland and Burke, 1988). Hence, for correlation purposes, the comparison of downslope soils may provide better data than the comparison of summit soils. Furthermore, one can compare the entire data for the catena with age or environment by calculating catena weighted-mean PDI values. This is done by multiplying the PDI value at each soil site by the fraction of the catena slope represented by that soil and summing the values for each catena. Birkeland and others (1991) have done this for a number of moraines in the Western United States (Fig. 5-6) and found that there is a rather tight grouping of values for the 20-ka moraines, and a spread of data points for the 140-ka moraines.

One major geomorphic problem with the PDI values (individual and catena weighted mean) described above is that, in places, the values are high even though the moraine might have undergone long-term erosion. Meierding (1984), for example, has constructed cross sections of 20- and 140-ka moraines (Fig. 5-7) that suggest considerable erosional lowering of the older moraines. The soil data described above fit with this erosional story if either (1) erosional adjustment (that is, moraine flattening) takes place relatively soon after moraine deposition and soil formation encompasses a subsequent longer time interval; (2) slower erosion upslope keeps the soils there poorly developed and the combination of continual deposition and soil formation downslope results in overthickened cumulic profiles; or (3) erosion and the resulting poorly developed soils upslope grade downslope to relatively stable, well-developed soils, and the materials eroded from upslope bypass the downslope sites.

Soil catenas in progressively more arid environments can be characterized by lateral translocation of only the more soluble compounds—the salts. For cold deserts, Glazovskaya (1968) noted the presence of the more soluble salts in soils at progressively lower parts of the landscape. Thus, $CaCO_3$ could exist in all the catena soils, but in the downslope direction, the more soluble salts appear in order of increasing solubility. Midslope positions could accumulate gypsum and Na_2SO_4, and chlorides of Ca, Mg, and Na could appear in the lowest parts of the landscape. Salts could move laterally if sufficient throughflow water is present. Iron released by weathering—which can be rapid in saline environments—would stay at the site of release and, along with aluminum, probably not be translocated laterally. The most likely clay mineral would be montmorillonite, but conditions appropriate for formation of fibrous clays such as palygorskite and sepolite also could exist.

At the wetter end of an arid-land transect, one might find carbonate leached from the upslope soils but accumulated in the downslope soils. In drier climates, however, carbonate would accumulate at all sites.

A soil catena formed from late Pleistocene to early Holocene loess in eastern Colorado (MAP 40; MAT 11) illustrates the conditions under which carbonate has accumulated in catena soils (Honeycutt and others, 1990a, 1990b).

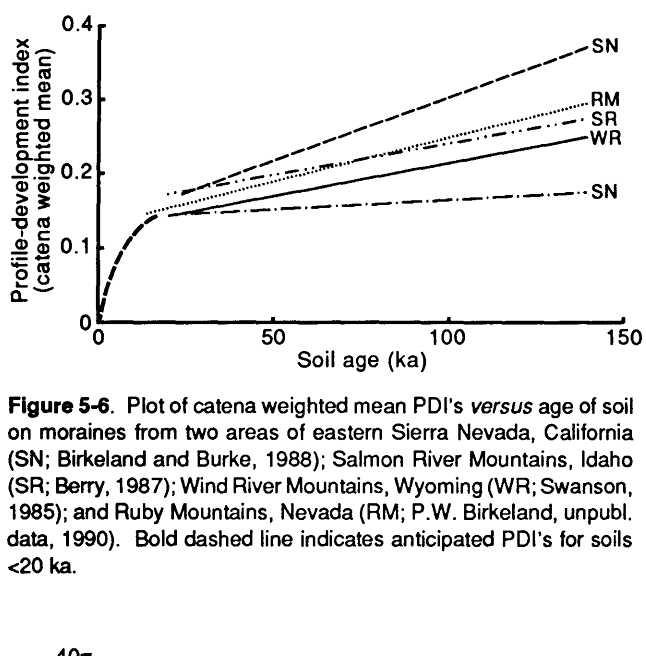


Figure 5-6. Plot of catena weighted mean PDI's *versus* age of soil on moraines from two areas of eastern Sierra Nevada, California (SN; Birkeland and Burke, 1988); Salmon River Mountains, Idaho (SR; Berry, 1987); Wind River Mountains, Wyoming (WR; Swanson, 1985); and Ruby Mountains, Nevada (RM; P.W. Birkeland, unpubl. data, 1990). Bold dashed line indicates anticipated PDI's for soils <20 ka.

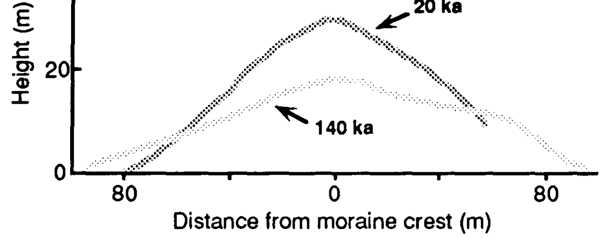


Figure 5-7. Typical cross-sectional profiles of Pinedale (*ca.* 20 ka) and Bull Lake (*ca.* 140 ka) lateral moraines in the upper Colorado River basin (from Meierding, 1984). Each profile represents an average of 10 profiles surveyed in the field. Similar profile *versus* age relations are seen at many localities in the Western United States.

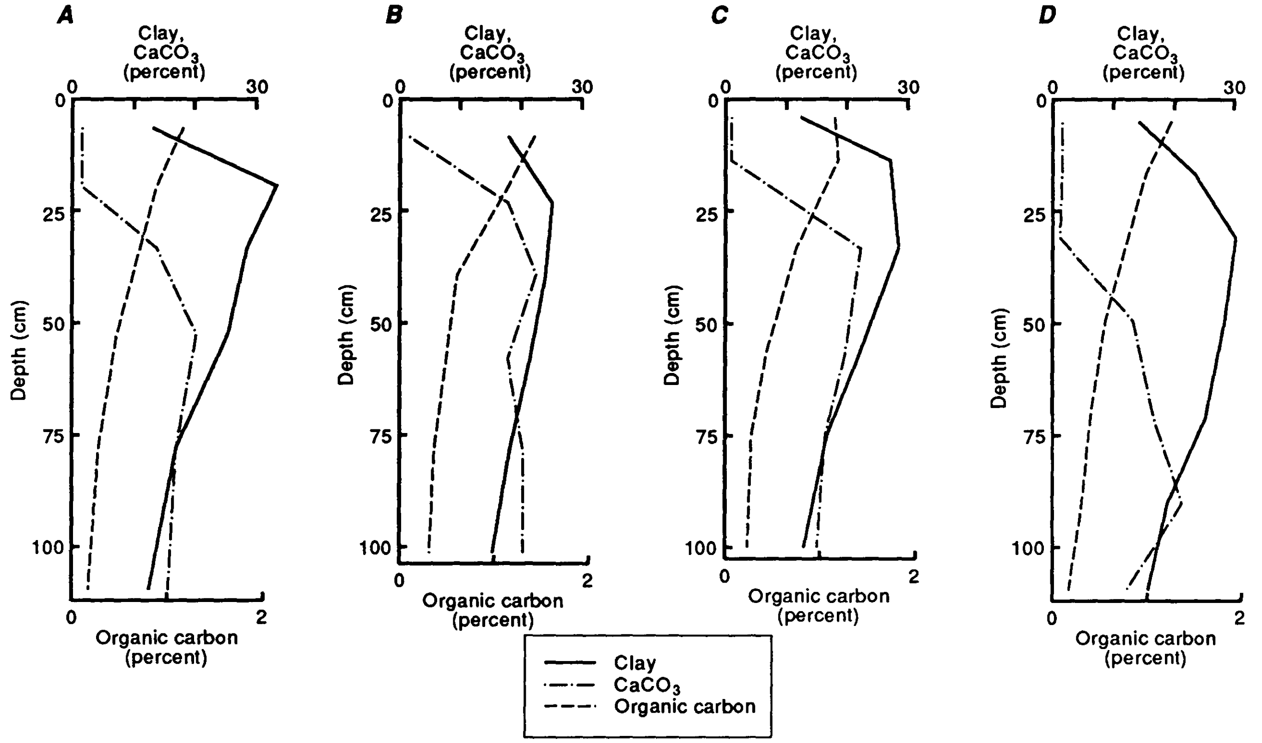


Figure 5-8. Vertical distribution of organic carbon, clay, and carbonate in a catena formed from loess in eastern Colorado (from Honeycutt and others, 1990a, 1990b). ***A.*** Summit. ***B.*** Shoulder. ***C.*** Upper backslope. ***D.*** Footslope.

The vertical distribution of organic carbon is similar at all sites (Fig. 5-8), suggesting little recent downslope movement of surface material. The clay distribution shows a slightly greater clay bulge at the footslope site, due either to greater clay production there or to the slow accumulation of clay at the footslope that was probably derived from upslope sites. The carbonate distribution curves are similar at all sites, but the carbonate bulge is at greater depth at the footslope site. Here, perhaps the microclimate at the footslope is slightly wetter and carbonate is leached to greater depths.

Birkeland (unpubl. data, 1990) has studied a series of synthetic (Fig. 5-2) river-terrace-scarp catenas near Las Cruces, New Mexico (MAP 20; MAT 17), where Gile and others (1981) conducted their classic soil studies. Catenas of three ages are being studied: less than 10 ka, *ca.* 20 ka, and *ca.* 75 ka. There is a change in the stages of carbonate morphology with time, from stages I-II on the youngest catenas to stages III-IV on the oldest catenas (Fig. 5-9). There are many similarities between the soil on the terrace surface (summit) and those on the terrace riser (footslope and backslope) of the same age. Carbonate content correlates well with the trends seen in carbonate morphology. Hence, in both this study and in the eastern Colorado study of Honeycutt and others (1990a, 1990b), soil formation seems to be dominant over erosion in the time since the catena slope was formed, so that somewhat similar profiles are seen at all sites in the catena.

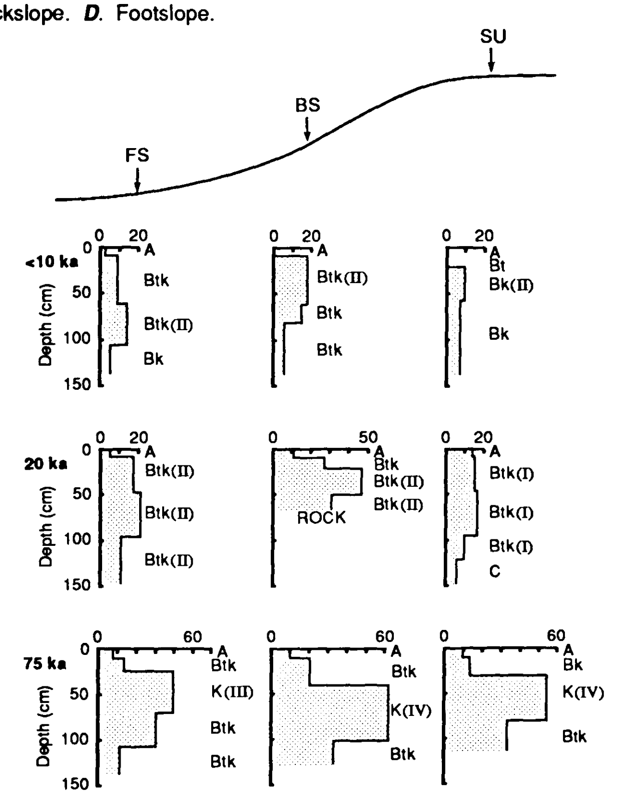


Figure 5-9. Synthetic-catena data for soil-horizon $CaCO_3$ content *versus* depth (from Birkeland, unpubl. data, 1990) for soils on river-terrace scarps of different ages along Arroyo Angostura, New Mexico (see map of Seager and Hawley, 1973). Horizontal axes are % $CaCO_3$, and maximum stage of carbonate morphology is given in parentheses.

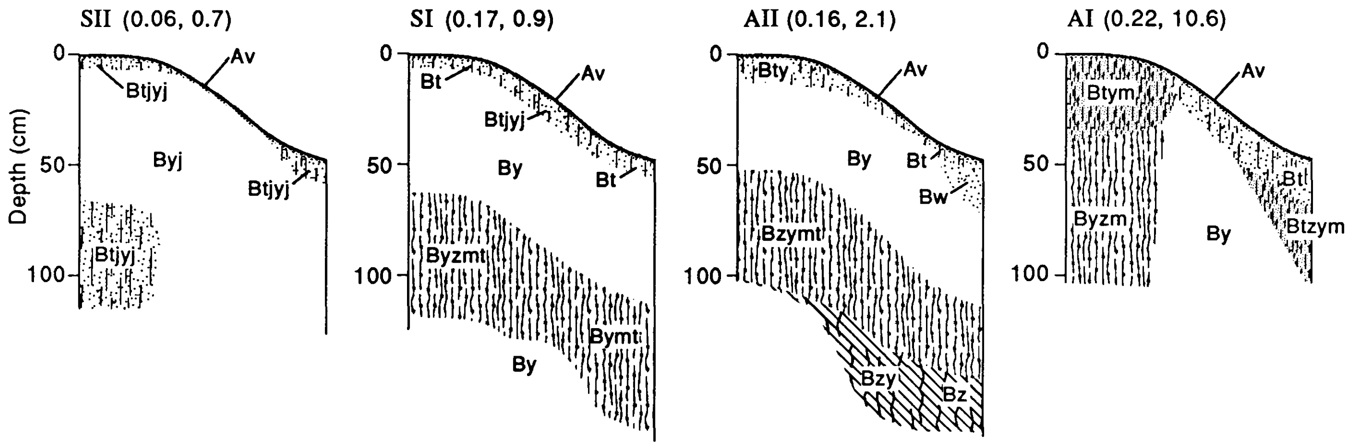


Figure 5-10. Soil catenas formed from river-terrace scarps in southern Israel (Birkeland, unpubl. data, 1990). These are not synthetic catenas, and the summit soils are found in the positions depicted here. The age sequence, from youngest to oldest, is SII, SI, AII, and AI. The first number in parentheses is the catena weighted-mean PDI, and the second number is the horizon PDI's summed to the base of the clay-accumulation horizon.

Soil catenas formed on river-terrace scarps have been described in the hyperarid climate (MAP 3; MAT 25) of southern Israel (Birkeland, 1986). The youngest catena is middle Holocene (SII, Fig. 5-10) and is characterized by weak Bt and By horizons. The next older catena (SI) is early Holocene to latest Pleistocene, has a thin Bt/By/Byzmt profile at most sites, and contrasts strongly with the younger catena, SII. The progressively older catenas (AII—younger late Pleistocene; AI—older late Pleistocene or middle Pleistocene) have thicker Bt horizons, and salts have accumulated in some Bt horizons. The PDI's for the catenas (weighted mean) display a good trend between the two youngest catenas, but values for the three oldest catenas are quite similar (Fig. 5-10). The best trends are for horizon PDI data summed to the base of the clay accumulation horizon. Depth plots of electrical conductivity, a surrogate for total salt content, also indicate similar values for the three older catenas but much less for the youngest catena. The only major trend with slope position is that the Bt thickens from the backslope to the footslope position in catena AI (the oldest catena), suggesting slight erosion and deposition, and the base of the Bt at the footslope is cemented by salts. Other than that, there seems to have been little redistribution of materials within the catenas by either geomorphic or pedologic processes in this hyperarid climate.

CONCLUDING REMARKS ABOUT CATENAS

The catena relations described above and others reviewed in Yaalon (1975) and Birkeland (1984) allow one to estimate soil ages for discrete portions of slopes. For example, a strongly developed soil with a red Bt horizon on a colluvial mountain slope in the Western United States suggests stability for about 100,000 yr. One could go further and map drainage basins, and assign general ages to various parts of the basin. Well-developed soils would indicate stable sites, whereas poorly developed soils indicate sites of active erosion or deposition. This really is an application of the K-cycle model of Butler (1959), where an attempt was made to place slope morphology in a time framework. The K-cycle model starts with a simple slope, such as the ones mentioned above. For a variety of reasons, the steeper, upper part of the slope might be unstable and erode fairly rapidly; this material can be deposited in the footslope-toeslope positions. If geomorphic stability follows, soil formation takes place along the entire slope. One K-cycle encompasses the erosional-depositional interval, as well as the soils that subsequently form over the entire slope. A subsequent period of instability, perhaps driven by climatic change, could initiate another K cycle. The aerial position and extent of both the erosion and deposition zones can change with time. The end result could be a mosaic of soils in which erosion keeps soils relatively poorly developed in the upper parts of the slopes and episodic deposition yields a sequence of buried soils in the lower parts of the slopes. Some areas of the slope could be beyond the areas of maximum erosion and deposition. If, instead of the periodicity of K cycles, erosion and deposition are more gradual through time, the soil might also reflect this. For example, instead of buried soils, one would see overthickened, cumulic profiles.

These same relations between soil and slope can be used to estimate the times of single or multiple faulting events (see Chapter VI). A soil on fault-scarp colluvium provides a minimum estimate on the time since the last faulting event, whereas buried soils in the colluvium put time limits on the hiatuses between faulting events.

In addition to the model presented earlier for clay minerals (Fig. 5-3), two other generalized models come from this work. One is for those areas in which the catena landforms are closely spaced in time and there is an accumulation of certain pedological constituents (*e.g.*, clay or Fe(d)) downslope. In this case, the relatively better developed soils in the downslope position of a particular-age landform may be similar to the summit soils of the next older landform, and continuing on, the downslope soil of the latter landform might be similar to the summit soil of the next still older landform.

The other model depicts soil catenas along a precipitation gradient in which all soils become more acid and downslope soils become more reducing at progressively higher precipitations, a model which extends that of Tonkin and others (1977). At low precipitation, similar gypsic soils are found at all slope positions, such as those discussed in southern Israel (Fig. 5-11). At slightly higher precipitation, gypsum does not accumulate but carbonate does, and the soils can be similar at all slope positions (*e.g.*, the New Mexico area). At still greater precipitation, calcic soils may only form at the downslope sites. The next step is to those precipitation levels where soils are leached, no carbonate accumulates, and accumulation of moisture in toeslope sites results in mottling. With even greater moisture and low enough pH, leached and oxidized soils may grade downslope to reduced soils, and finally reduced soils might form across the whole landscape. The outcome of this model is that the morphological contrasts, and perhaps geochemical ones, from site to site in the catena are at a maximum at the intermediate precipitation levels, and decrease toward both lower and higher precipitations to reach minimum contrast at the lowest and highest precipitation.

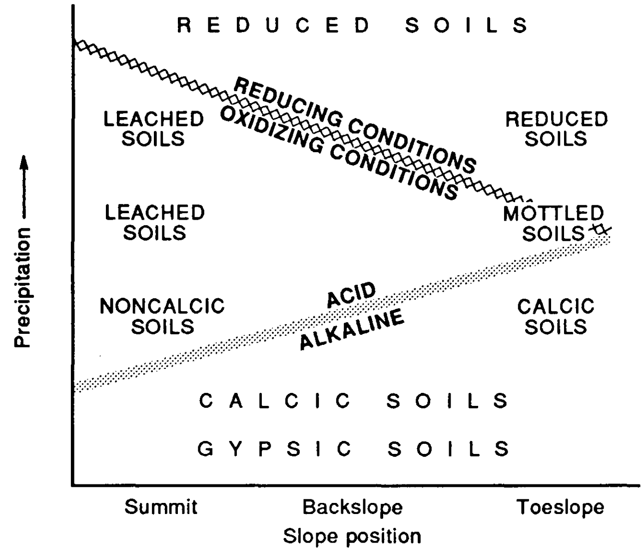


Figure 5-11. Schematic diagram depicting type of soil *versus* catena position along a precipitation gradient in which pH becomes progressively lower and reducing conditions become more extensive (laterally) at higher precipitations (modified from Tonkin and others, 1977).

CHAPTER VI
CASE STUDIES OF SOILS IN APPLIED GEOLOGY

INTRODUCTION

This chapter presents a number of examples of how soils have been used in applied geology. Soils have been used to reconstruct climate on the basis of overall soil morphology, Bt horizon properties, and the position of $CaCO_3$ accumulation and the morphology of K horizons. In addition, we discuss ancient paleosols—their recognition and interpretation. Perhaps the most common use of soils in geology is as a tool for surficial geologic mapping; in this case, we review soil studies in the Beaver basin of central Utah. Finally, we will discuss some of the uses of soils as they apply to studies of Quaternary tectonics, specifically the timing of faulting events, the use of soils for deciphering structural relations, and how soil-age estimates might be used to determine slip rates.

RECONSTRUCTION OF PAST CLIMATES FROM PEDOLOGIC DATA

Morphology and various other properties of soils can be used to infer past climates. These inferences are important to Quaternary research because, in many places, the soils represent hiatuses in the depositional record and, thus, may be the only record left of certain time intervals. Before paleoclimatic interpretations can be made, however, one has to be sure that the observed feature was truly imparted on the soil by a past climate and not by other factors of soil formation. For example, it is commonly recognized that the effects of long intervals of soil formation can give the same pedologic result as those of some climatic changes.

Soil properties vary in their usefulness as tools for paleoclimatic interpretation. If the soil has remained at the surface since the climatic change, the property or properties used to decipher changes in climate must have been resistant enough to persist in the soil, and not be altered entirely during subsequent pedogenesis. Obviously, those soil properties that alter readily with changing environmental conditions, such as pH and organic-matter content, cannot be used as indicators of past conditions. The same is true for paleoclimatic interpretations of buried soils. In this case, the properties imparted to the soil during pedogenesis must be resistant to subsequent diagenetic alteration to be useful in paleoclimatic reconstruction. Morphological features can be ranked by their persistence under changing environmental conditions or burial (Table 6-1). For example, those horizons or features ranked as persistent or relatively persistent are more apt to carry a legacy of former environmental conditions. Success in deciphering paleoclimatic conditions from soils evidence will also depend partly on the field area one chooses. The most promising areas are those of rather sharp transition from one climatic type to another because there the soils are most sensitive to changing environmental conditions.

The generalized model of Machette (1985a) is good for portraying the effects of climatic change. Although constructed to demonstrate the influence of climatic change on pedogenic carbonate in soils, the model is relevant to a wider variety of environments and materials (Fig. 6-1). Two general fields are defined in his model. The influx-limited field is defined by areas where there is sufficient moisture to translocate material into the soil in either dissolved or solid forms, so potentially most of the materials deposited on the surface are moved into the soil. In contrast, in the moisture-limited field, there is insufficient moisture, relative to the influx rate, to move all of the materials deposited at the surface into the soil; subsequent wind or surface erosion can move the materials to other sites before pedogenic processes move them into the soil.

Several examples can be used to show how the model works (Fig. 6-1). In area A, salts are leached from the soils during glacials and accumulate during interglacials. Atmospheric dust influx, derived from deposits in pluvial lake basins that are dry during interglacials might follow the same trends. However, if dust influx is greatest during glacials (*i.e.*, loess from active glacial-outwash flood plains) the

Table 6-1.—Relative persistence of soil horizons and properties as possible indicators of former pedologic conditions [Modified, in part, from Yaalon (1971b)]

LEAST PERSISTENT	RELATIVELY PERSISTENT	PERSISTENT
A horizon	E horizon	Bt, Bq horizon
By, Bz horizons	Bw horizon	K horizon
pH, base saturation	Bk horizon	Fe content
	Fe content	Carbonate content
	Carbonate content	Clay mineralogy
	Clay mineralogy	

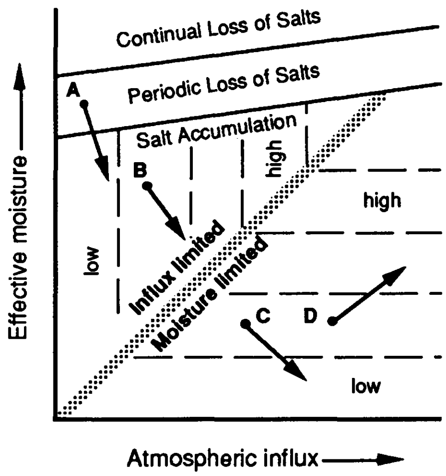


Figure 6-1. Simplified model of relative pedogenic accumulation rates (dashed vertical and horizontal lines) as a function of the amounts of effective moisture and atmospheric influx. Atmospheric influx can be salts (calcium carbonate, gypsum, halite) or solids (silt and clay). Arrows depict the direction of possible changes in pedogenic accumulation rates, with the end of the arrows depicting conditions under full-glacial climate and the point of the arrows depicting conditions under interglacial climate. (Modified from Machette, 1985a).

direction of the arrow would be reversed. In area B, accumulation takes place at all times, but the rate will vary from glacial to interglacial. In area C, the rate of salt accumulation could decrease during interglacials because there is insufficient moisture to dissolve and translocate it to depth in the soil. Dust influx could behave in the same manner. Finally, in area D, for reasons of bedrock outcrop patterns, dust trajectories, and local salt or dust sources (such as near Salt Lake City), pedogenic accumulation rates are higher during interglacials.

Overall Soil Morphology

Some soils in a region, either at the surface or buried, might have a morphology quite different from currently forming soils in the region. In such cases, it may be possible to reconstruct the past climate by comparing the present-day climate and morphology of soils in the region with the climate and morphology of the soil or soils in question. The best way to undertake this kind of study is with soil climosequences, such as a basin transect in the Western United States. In many places, A/E/B profiles are typical of these forested mountain soils, and the E horizons are not expected in the grassland basin soils. However, in South Park, Colorado, the basin soils have an E horizon under a short-grass prairie that is several kilometers from the nearest forest. The E horizon could mean that forest vegetation occupied parts of the valley floor in the past and that the subsequent organic-matter production in the short-grass prairie has not been sufficiently rapid to obliterate the E horizon.

Bt-Horizon Properties

The position of the Bt horizon, its color, and the amount of clay are factors that might contribute to an understanding of past climates. Unambiguous interpretations are hard to make, however.

The position of the Bt horizon should relate to the climate under which the soil formed. It is commonly reported that aridic soils are shallow (thin) and that soils in progressively wetter climates are progressively thicker; the Bt-horizon position would parallel these trends. One might be able to relate the position of the Bt horizon to the water-holding capacity of the soil and to water movement within the soil for various climatic regions.

A surface soil that is redder than younger surface soils in the area might be used as an indication of a former warmer climate. This can only be done, however, with equivalent parent materials. Moreover, the duration of soil formation has to be known with certainty so that the time factor can be ruled out as a cause of the redness. If it can be demonstrated that the red soil formed over a time span similar to that over which the younger non-red soils formed, a temperature change might well have occurred (Schwertmann and others, 1982).

The amount of clay present in a soil can be used to indicate past climatic change, if the duration of soil formation is precisely known. Here, if the clay content of an old buried soil is greater than that in a soil formed under the present climate, and if both soils formed over a similar time span, it can be concluded that the older soil formed in an environment conducive to a more rapid accumulation of clay. To increase the rate of clay accumulation, one might postulate an increase in precipitation, an increase in temperature—both of which could stimulate faster weathering—or greater eolian dust influx.

The above approach to paleoclimate can be demonstrated with a field example in the Western United States. Hunt and Sokoloff (1950) remarked that strongly developed pre-Wisconsin surface soils of the Rocky Mountain region may have developed in the same time span as the weakly developed post-Wisconsin soils, but under different climatic conditions. This may well be true, but there is no known climatic region in the United States in which post-Wisconsin soils (*ca.* 10-15 ka) are anywhere as strongly developed as the pre-Wisconsin soils (*ca.* 120-140 ka or older) they describe.

Changing depths of soil-moisture penetration during glacial and interglacial periods may be indicated by the distribution of clay minerals with depth. In the Yukon Territory, for example, the pedogenic alteration of primary clay minerals in older soils extends to depths greater than that suggested by water movement in the present-day climate (Foscolas and others, 1977) and, thus, suggests greater soil moisture in the past. Karlstrom (1988) used a similar line of reasoning to propose climatic change in the Rocky Mountains of the United States and Canada.

Position and Morphology of $CaCO_3$ Accumulation

Because the presence or absence of $CaCO_3$ in a soil is directly related to soil-water movement, and hence to climate, the position and morphological features of $CaCO_3$ accumulations may provide insight into past climates. The position of the top of the horizon of $CaCO_3$ accumulation is related to the regional climate, as long as the water-holding capacity of the soil is taken into account. Arkley (1963) found that the tops of many of the Bk or K horizons plot close to the sharp break in the calculated soil-water-movement curves that represent a rapid decrease in water movement with depth (Fig. 6-2). During a change to a drier climate, the curve shifts to the left and all accumulation products (clay, carbonate, *etc.*) are concentrated higher in the profile. In contrast, an increasingly moist climate shifts the curve to the right, with carbonate either leached entirely from the system or shifted to a greater depth. Birkeland has seen this latter case in one locality in Wyoming, where the parent material is a well-drained gravel of a high river terrace. About 20 cm of unweathered material (Cu horizon) separates the base of the Bt horizon from the top of the Bk horizon. One interpretation is that the top of the Bk was 20 cm higher in the past and that weathering of iron-bearing minerals was inhibited in the carbonate environment. A subsequent climatic change resulted in deeper water penetration and shifted the top of the Bk horizon to greater depths, but the change has been recent enough so that the silicate grains in the former Bk horizon have not altered to form either a Cox or a Bw horizon.

Other evidence for climatic change is a Btk horizon in which clay and carbonate accumulations have alternated at a particular depth interval through time. This would happen if the wetting curves alternated between A and B in Figure 6-2. Reheis (1987b, 1987c) found evidence for this in southern Montana (Fig. 6-3), where the depths of pedogenic clay and carbonate accumulations overlap for soils of all ages. In thin section, overlapping clay and carbonate films are present, and the number of overlapping pairs of films is greater on older deposits. In her analysis of possible climatic changes during the Quaternary, she suggests that glacial times probably were times of greater effective soil moisture, and this would be indicated by a soil-moisture wetting curve to the right of the one shown in Figure 6-3. Hence, she proposes this interpretation: glacial times, with the curve

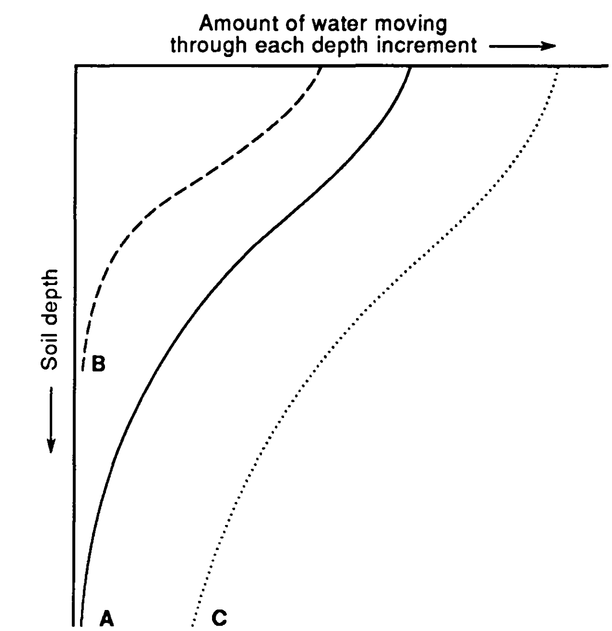


Figure 6-2. Schematic diagram showing the movement of water with depth, according to the model of Arkley (1963). Curves: **A** is any one time and the top of the Bk horizon is close to where the curve becomes asymptotic with the vertical axis; **B** is after a change to a drier climate, in which case the top of the subsequent Bk horizon is much closer to the surface, and all properties formed during curve **A** are preserved; and **C** is after a change to a wetter climate, which would result in eventual removal of the more soluble soil constituents and a greater depth of clay accumulation.

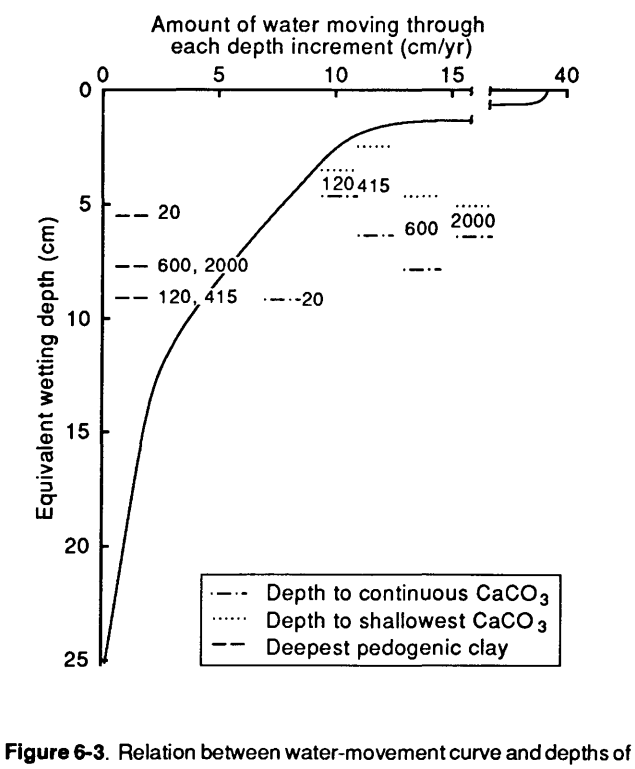


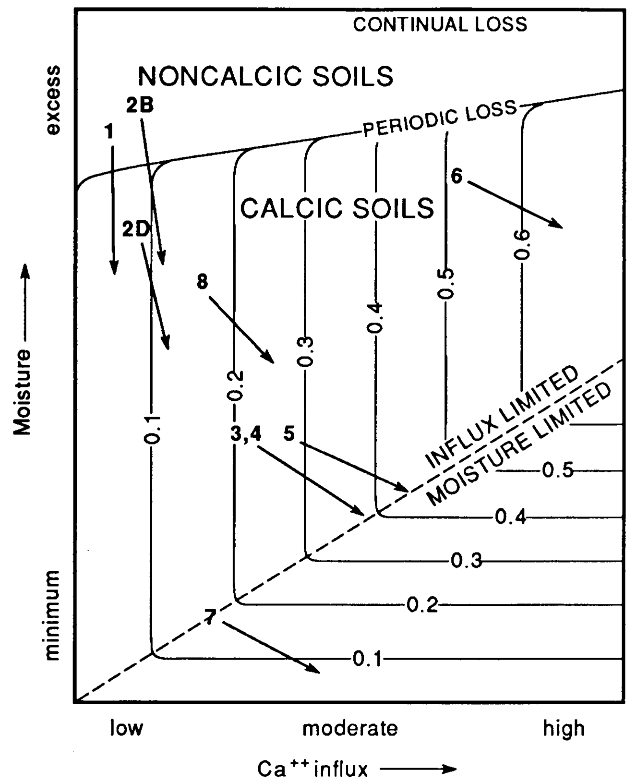
Figure 6-3. Relation between water-movement curve and depths of pedogenic clay and carbonate for soils in a transitional area between the mountains and basin, southern Montana (from Reheis, 1987b). Equivalent wetting depth is the depth of soil expressed in centimeters of available water-holding capacity (see Arkley, 1963). Numbers adjacent to depth symbols are the estimated soil ages in ka.

shifted to the right, were marked by greater water penetration and clay translocation; and interglacial times were marked by shallower water penetration and carbonate films deposited on previously deposited clay films.

Machette (1985a) modeled the rate of carbonate accumulation under changing climatic conditions for various localities in the Western United States (Fig. 6-4). Estimates were made not only on the direction of the rate with climatic change, but also for the amounts of accumulation. Soils that plot in the influx-limited portion of the model have greater rates of carbonate influx during the drier interpluvials owing to increased dust flux. Some of this carbonate could be derived from dry lake beds or limestone terrain that has a less dense cover of vegetation during interpluvials. Data from one soil, which plots in the moisture-limited field, has a decreased rate of carbonate influx during interpluvials; the explanation is that moisture is insufficient during interpluvials to move all of the relatively abundant $CaCO_3$ into the soil.

McFadden and Tinsley (1985) and Mayer and others (1988) modeled the important parameters of calcic soils (soil CO_2, dust flux, water movement, *etc.*) to determine the depth of carbonate accumulation in aridic climates. In one case, they mimicked the distributions in desert soils that have gone through the late Pleistocene (relatively wet climate and deep carbonate accumulation) to Holocene (relatively dry climate and shallow accumulation) climatic change. In another case, however, the modeled carbonate curve indicates much less carbonate than is actually present (Fig. 6-5). The model may have underestimated either (or both) the past carbonate-influx rates and the amount of effective moisture.

A final example relates to the soil-forming intervals of Morrison (1978) and alerts one to the problem of differentiating the time and climate factors in soil studies. A soil-forming interval is a period of time during which soil formation was relatively rapid and is bracketed by times during which soil formation was relatively slow. Interglacials or interstadials were said to be the times most conducive to soil formation. An example of soil-forming intervals involves the time to form the Promontory Soil—a prominent soil that was buried by the last major rise in Lake Bonneville. The Promontory commonly is a calcic soil with Bt and Bk or K horizons, with a maximum carbonate morphology of stage III. Morrison and Frye (1965) interpreted this soil as having formed in less than 10,000 yr; this would require unusually rapid formation under a former different climate. However, recent stratigraphic and chronometric work suggests that the Promontory and other similar soils in the Bonneville basin took a much longer time to form (*ca.* 100,000 yr), using carbonate-accumulation rates similar to those of the Holocene (Scott and others, 1983). This example does not negate the influence of past climates and climatic changes on soil formation but points out the difficulty of separating the effect of climatic change from the effect of time, especially when one lacks independent age control.



LOCATIONS			
1	Buena Vista, Colo.	5	Las Cruces, N. M.
2	Boulder-Denver, Colo.	6	Roswell-Carlsbad, N. M.
3	Albuquerque, N. M.	7	Vidal Junction, Calif.
4	San Acacia, N. M.	8	Beaver, Utah

Figure 6-4. Long-term carbonate accumulation rates shown by vertical and horizontal lines (in g of $CaCO_3/cm^2/10^3 yr$) under varying conditions of moisture and Ca^{2+} influx (from Machette, 1985a). Measured accumulation rates for the chronosequences during pluvial episodes are plotted at the end of the arrows; the points of the arrows show inferred rates and conditions during interpluvial episodes. See text for explanation of influx-limited and moisture-limited parts of the diagram.

Ancient Paleosols—Recognition and Interpretation

Buried soils (paleosols) are reported in sedimentary deposits and rocks of many ages (Retallack, 1983; Bown and Kraus, 1987; Kraus, 1987; Wright, 1987; Reinhardt and Sigleo, 1988). Paleosols are used to recognize unconformities, estimate the duration of time represented by an unconformity, and estimate the paleoenvironment at the time the unconformity was created.

Recognizing pre-Quaternary paleosols is not easy. Biological traces (fossil tracks, burrows, root casts, *etc.*) in rocks indicate that a land surface existed, and, therefore, some sort of a soil may have formed. Some pedological features commonly recognized in contemporary soils, however, are not commonly observed or preserved in paleosols. These include an organic-matter enriched A horizon, structure,

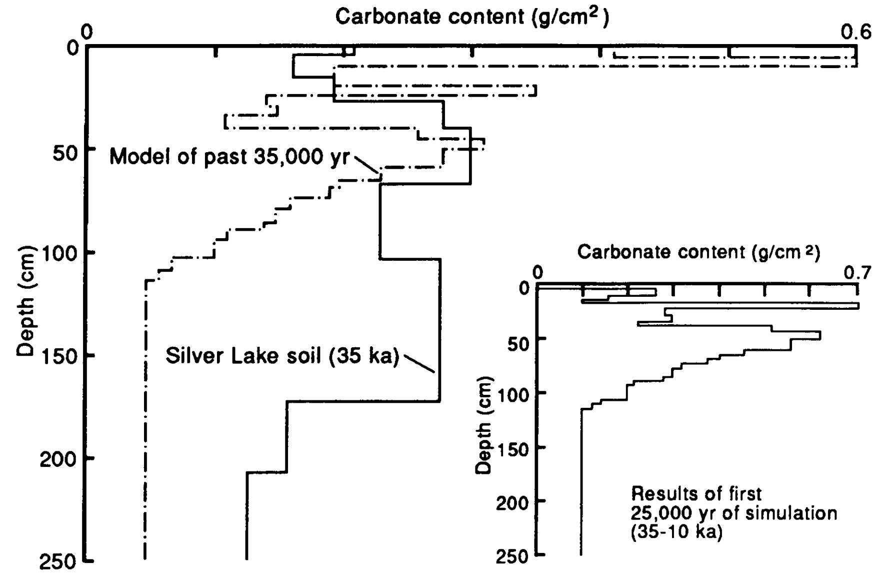


Figure 6-5. Plot of carbonate distribution for latest Pleistocene soil from Silver Lake, California, compared to modeled carbonate distribution (from Mayer and others, 1988). The model incorporates two distinct climates; a wet one for the period 35,000-10,000 yr (see inset; note that surface horizon is carbonate free) and a dry one for the past 10,000 yr, during which time carbonate accumulated at the surface.

vertical variation in pedogenic coloration, clay films, and carbonate morphologies common in aridic soils. The main problem is that many features that have been ascribed to possible pedogenesis could also have formed through burial diagenesis, and sorting out the effects of these two processes is difficult and controversial (Patterson, 1990; Patterson and others, 1990).

Assuming that one or more of these features is indicative of a soil, duration of soil formation can be estimated. For example, the duration can be estimated by comparison with clay accumulation curves (Fig. 4-7) or carbonate-accumulation stages (Fig. 4-6). Extraction of specific pedogenic elements, such as Fe(d), might not work for comparison because pedogenic Fe could be too altered following burial to survive and allow for an unambiguous pedological interpretation. Additionally, total chemical analysis may not give useful age trends, because the best trends in these data seem to require highly leaching environments over a long duration of time (>100,000 yr) for pedological trends to overprint the parent-material signal. Prior to undertaking this type of analysis, however, one should review in detail the total-chemistry data for soil chronosequences in a variety of environments (as described in U.S. Geological Survey Bulletins 1589 and 1590) to look for trends pertinent to their study.

The other common use of paleosols is to infer paleoenvironment. A paleoenvironment might be broadly inferred from the kind of soluble salt in the profile as well as its depth relation, and perhaps by the depth of clay migration in the soils, as long as diagenesis can be taken into account or ruled out. Although classification of paleosols under the rules of soil taxonomy (Soil Survey Staff, 1975) has been proposed to aid in paleoenvironmental reconstruction (Retallack, 1983), it appears that too many of the key classification criteria have been lost on burial diagenesis for this approach to be used in any detail. Overall morphology is perhaps the best way to estimate past climate at the time of soil formation, but this is not easily done.

SOILS AS A TOOL FOR SURFICIAL GEOLOGIC MAPPING

Soil Studies in the Beaver Basin of Central Utah

The Beaver basin of central Utah is a classic intermontane basin that formed in the early to middle Miocene and was subsequently filled by terrestrial and lacustrine sediment during the late Miocene and early Quaternary. This basin was host to a perennial lake (Lake Beaver; Machette, 1985b) which occupied the basin in the late Cenozoic and into which fell as many as eight distinct rhyolitic and basaltic volcanic ashes of local and distant origin. Overtopping of a low basaltic sill in the southwestern part of the basin in the early(?) Quaternary allowed Lake Beaver to drain, and the Beaver River and its main tributaries (Indian Creek and North Creek) to flow unimpeded to a significantly lower base level at Milford, Utah, in the eastern Great Basin.

In the succeeding time, these streams cut down from their pre-incision level (Table Grounds surface; Fig. 6-6) through the ancient floor of the lake basin and into the relatively soft sediment, thereby causing an inversion of topography in the ancient basin. The initial phase of downcutting resulted in a widespread, gravel-covered pediment surface named Last

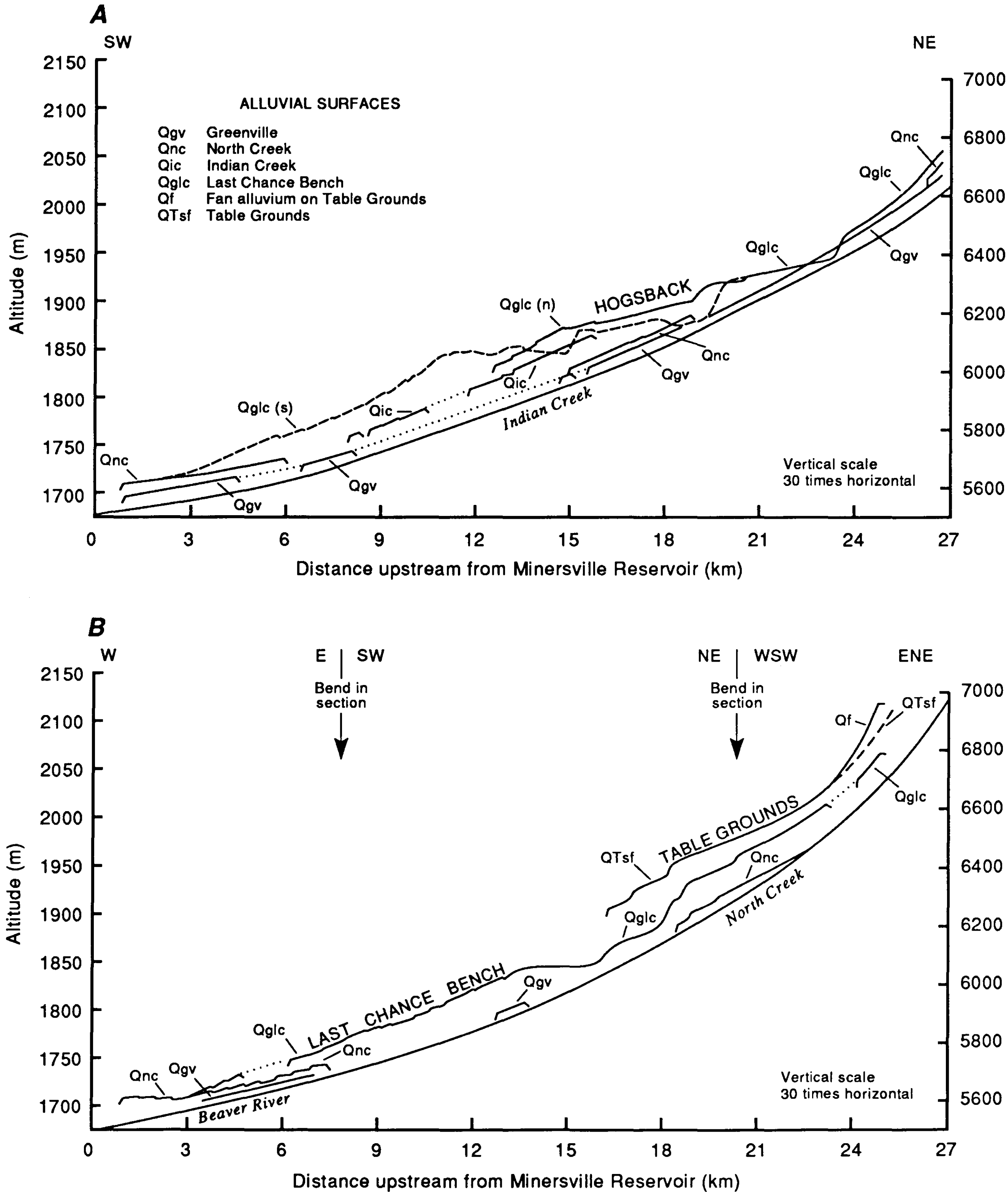
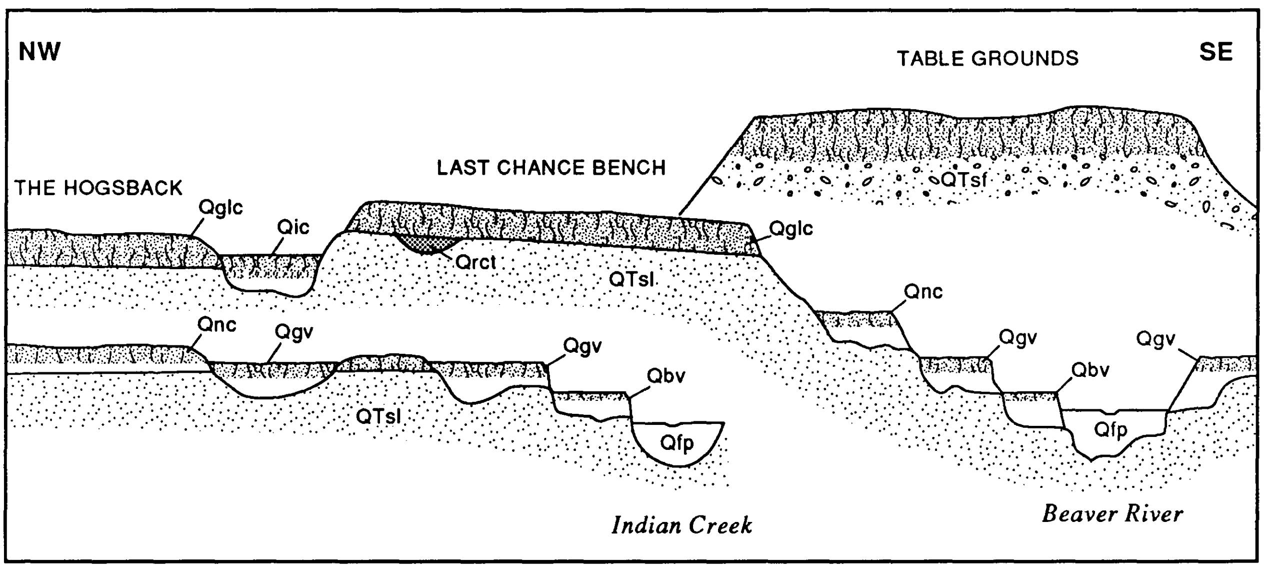


Figure 6-6. Topographic profiles of deformed alluvial terraces along drainages in the Beaver basin, Utah (Machette, unpubl. data, 1985). ***A***. Profiles along Indian Creek, an intermittent stream in the northern part of the basin. Dashed line indicates the Qglc surface south of the creek. ***B***. Profiles along the Beaver River and North Creek, a perennial stream that is a north fork of the Beaver River in the east-central part of the basin.

Chance Bench. Occasional backfilling events (probably climatically related) during the subsequent erosion of the basin have created distinct stream terraces and alluvial-fan surfaces, only parts of which are preserved today. To complicate matters, the basin has been subject to extensive deformation, both in the form of a north-trending axial antiform and as a result of basin-margin faults that form the transition zone between the Colorado Plateaus province to the east and the Basin and Range province to the west. The mapper of surficial geology in this and similar basins may have one or more of the following objectives:

1. To map and correlate stream-terrace, fan-complex, and other surficial deposits on a firm basis throughout the basin. Elevation above stream level is not a reliable criteria in settings where drainage changes have occurred or where tectonic deformation is suspected.
2. To estimate the time of breaching of the basin, and thus, the maximum age of the erosional incision of the basin.
3. To estimate, in at least relative terms, the ages of the fan and terrace remnants in the basin.
4. To make correlations with glacial/interglacial episodes and establish a model for stream incision and alluvial deposition that may be climatically driven.
5. To estimate rates of deformation (antiformal, in this case) and fault slip, and to estimate the timing of individual faulting events (typically the youngest event).

The Beaver basin lends itself to the type of mapping-based studies described above (Machette, 1983; Machette and others, 1983; Machette and Steven, 1983) and, in fact, each of these objectives was part of their studies. The field description and laboratory characterization of selected soils on alluvial units, which were differentiated and mapped using conventional methods (aerial photography, interpretation of soil-survey data, and field inspections), allowed Machette (1982, 1985b) to estimate the general ages of soils on the various alluvial units (Fig. 6-7) and to generally



SYMBOL	DESCRIPTION	AGE (YEARS)
Qfp	Holocene alluvium	Holocene (less than 10,000)
Qbv	Alluvium of Beaver	Latest Pleistocene (12,000 - 15,000)
Qgv	Alluvium of Greenville	Late Pleistocene (120,000 - 140,000)
Qnc	Alluvium of North Creek	Middle Pleistocene (250,000)
Qic	Alluvium of Indian Creek	Middle Pleistocene (350,000 - 400,000)
Qglc	Gravel of Last Chance Bench	Middle Pleistocene (500,000)
Qrct	Tephra of Ranch Canyon	Middle Pleistocene (550,000)
QTsf	Basin fill, coarse	Early Pleistocene to Pliocene (0.75 - 5 Ma)
QTsl	Basin fill, fine	Early Pleistocene to Pliocene (0.75 - 5 Ma)
	Soils in alluvium	Pattern proportional to depth and degree of soil development

Figure 6-7. Schematic cross section of alluvial terrace levels, soil development, and age assignments (from Machette, 1985b).

confirm the physical correlation of isolated alluvial remnants (Fig. 6-6). The soils in the Beaver basin cross the pedocal/pedalfer boundary; that is, across the zones of carbonate accumulation and leaching. This boundary is present in late to middle Pleistocene soils at about 1,950 m and marks the present transition from sagebrush and short grass to piñon pine/juniper forest. However, this same zone is lower and basinward in latest Pleistocene and Holocene soils, which reflects a relatively high water table in these deposits. Conversely, the oldest soils in the basin are calcareous above the pedocal/pedalfer boundary, indicating that the boundary probably fluctuates through a altitudinal range of several hundred meters in response to climatic change.

The continual growth of an antiform (keystone-like anticline) in the heart of the Beaver basin and basin-margin faulting has caused the oldest surficial deposits to be extensively deformed; in some cases the gravel of Last Chance Bench is offset as much as 30 m along discrete faults, but the younger deposits show similar patterns, albeit lesser amounts of deformation. Topographic profiles were constructed in order to correlate these deformed deposits, particularly those that are isolated remnants (Fig. 6-6).

The time of breaching of the Beaver basin was constrained from three lines of evidence. The first is an age of 540 ka (see Machette, 1985b) from the rhyolitic tephra of Ranch Canyon (Fig. 6-7) that is present in drainage channels incised in the basin fill and which indicates exterior drainage for the basin. The second piece of evidence is the development of a calcic soil on the gravel of Last Chance Bench, which directly overlies the 540-ka ash. This soil is well developed, having stage III+ morphology (Table 6-2) and about 70 ± 5 g of $CaCO_3/cm^2$ of soil in the profile, and a uranium-trend age of >420 ka. The third line of evidence is somewhat subjective, but helpful. The soil on the Table Grounds surface, which represents the alluvial apron on the east side

Table 6-2.—Key soil data for major alluvial surfaces in the Beaver basin, Utah
[Symbols; t, thickness; *t, thickness in parentheses is entire clay accumulation zone (includes $CaCO_3$ engulfed Bt horzions); *Clay, net increase relative to horizon in parentheses; *$CaCO_3$, maximum content in <2mm fraction of calcic horizon. Location of soils shown in Machette, 1982. Modified from Machette, 1982, 1985b]

Soil	Bt Horizon			Bk/K Horizon			Total	Remarks
	*t (cm)	Maximum Color (Munsell)	*Clay	t (cm)	Maximum Stage	*$CaCO_3$	$CaCO_3$ (g/cm^2)	
SOIL IN ALLUVIUM OF BEAVER (LATEST PLEISTOCENE, 12-15 KA)								
No. 1	30	*5YR* 5/3 to 5/4 (d)	3 (A)	100	I	0.4	n.d.	High water table
No. 2	35	*7.5YR* 4/4 to 5/4 (d)	3 (A)	20-120	II-	0.7	n.d.	High water table
No. 3	25	*7.5YR* 4/4 (m)	4 (A)	110	I+	1.9	0.3	
Preferred Values	30	*7.5YR* 5/4 (d)	3 (A)	100	I+	2.0	<0.3	
SOIL IN ALLUVIUM OF GREENVILLE (LATE PLEISTOCENE, 130-140 KA)								
No. 4	17	*5YR* 5/3 to 5/4 (d)	20 (C)	56	III	32	11	
No. 5	31	*7.5YR* 5/4 (m)	29 (C)	53	III-	27	8	Thin profile; $CaCO_3$ lost
No. 6	45	*7.5YR* 5/5 (m)	16 (C)	52	III-	16	6	Leached; high water table
Preferred Values	30	*7.5YR* 5/4 (d)	22 (C)	54	III-	25	8-11	
SOIL IN ALLUVIUM OF NORTH CREEK (MIDDLE PLEISTOCENE, 250 KA)								
No. 7	60 (85)	*5YR* 5/6 (d)	25 (C)	118	III	47	25	Loess over gravel
No. 8	35 (68)	*5YR* 5/6 (d)	32 (C)	>85	III	15	>9	Thin; $CaCO_3$ lost
No. 9	47 (88)	*5YR* 6/6 (d)	25 (C)	115	III	40	38	Thick; U-trend age >240 ka
Preferred Values	51 (80)	*5YR* 5/6 (d)	27 (C)	106	III	44	33±5	
SOIL IN GRAVEL OF LAST CHANCE BENCH (MIDDLE PLEISTOCENE, 500 KA)								
No. 10	35 (107)	*7.5YR* 4/3 (d)	28 (C)	127	III+	59	78	Max. clay is in K horizon
No. 11	67 (132)	*5YR* 5/6 (d)	32 (C)	>133	III+	53	49	Base covered; min. $CaCO_3$
No. 12	18 (100)	*5YR* 4/4 (m)	24 (C)	132	III+	68	70	U-trend age >420 ka
Preferred Values	40 (100)	*5YR* 5/5 (d)	28 (C)	130	III+	60	74±4	
SOIL BELOW TABLE GROUNDS SURFACE (EARLY PLEISTOCENE, 750 KA)								
No. 13	none	n.d.	n.d.	>110	IV	50	>51	Eroded; covered base
No. 14	27	*7.5YR* 5/4 (m)	n.d.	>150	IV	65	66	High altitude; $CaCO_3$ lost
Preferred Values	stripped, consumed	*7.5YR* 5/4 (m)	n.d.	>150	IV	65	>51	$CaCO_3$ lost

of the Beaver basin prior to incision, is considerably better developed than that on Last Chance Bench. The soil on the Table Grounds has an eroded Bt horizon, and a thick stage IV morphology K horizon. This soil is above the general altitude of the pedocal/pedalfer boundary in the basin and, thus, its total $CaCO_3$ of >51 to 66 g/cm^2 probably represents occasional loses due to leaching. On the basis of these three lines of evidence, Machette (1985b) estimated an age of 750 ka for the Table Grounds surface and, thus, a minimum time of basin incision.

Soils are useful for making estimates of the ages of the fan and terrace remnants in the basin (Fig. 6-7). Data for both clay accumulation in the Bt horizons (Fig. 6-8*A*) and calcium carbonate in Bk/K horizons (Fig. 6-8*B*) show a general increase with age, but all of the measured parameters fall off in the upper age range (typically at or above 500 ka). These relations are due to loss of Bk and K horizons as alluvial surfaces are eroded, and to consumption of Bt and Bk horizons by upward accumulation of carbonate owing to the inherent increased water-holding capacity of soils with time. In order to estimate soil ages on the basis of some quantitative measure of soil formation, a calibration point is needed. If one assumes that the soil on Last Chance Bench started to form on the Bench at about 500 ka, then the long-term rate of carbonate accumulation is 0.14 g/cm^2/yr. Using this rate, Machette (1982, 1985b) estimated the duration of soil formation necessary to develop the soils on each of the major alluvial surfaces in the Beaver basin (Fig. 6-7 and Table 6-2). The apparent timing of stream aggradation and incision, and the magnitude of time between the last several major hydrologic cycles in the basin suggests that aggradation may have occurred during the transition from glacial to interglacial episodes, when large changes in streamflow, sediment supply, and climatic conditions probably occurred.

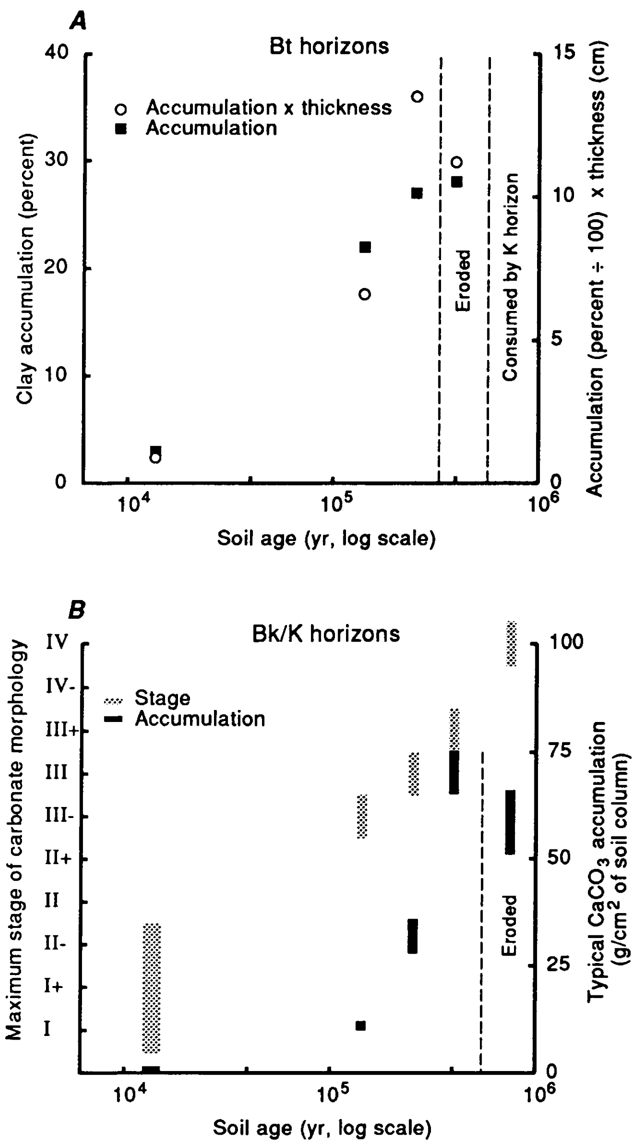


Figure 6-8. Plots of field and laboratory data for soils of about 15-ka to 750-ka age in the Beaver basin. ***A***. Clay accumulation (in percent and percent times thickness, which is a proxy for total clay content). ***B***. Calcium carbonate development (as maximum stage of morphology and total pedogenic accumulation).

The final goal of the Beaver basin studies was to estimate rates of deformation (antiformal, in this case) and fault slip, and to estimate the timing of individual faulting events (typically the youngest event). These subjects are discussed in the following section.

SOILS APPLIED TO STUDIES OF QUATERNARY FAULTING

Timing of Faulting Events

The development of soils represents periods of landscape stability—that is, times when the processes of erosion or deposition are outpaced by soil-forming processes. As such, soils can be used in tectonic studies to indicate periods of tectonic stability and, thus, estimate the timing of faulting events. In the simplest form, timing estimates are made from the differences in development between soils formed on the upthrown (stable) block, on the downdropped (buried) block, and on colluvium shed from the scarp during a discrete faulting event. For example, in Figure 6-9 some measure of soil development (PDI; g of $CaCO_3$ or clay) of the relict soil on the upthrown block is taken as 100%. Then, if the fault-scarp locality is a closed system, the sum of soil development for all of the buried soils and the surface soil should also be 100%. Using this relation and the age of the relict soil (here assumed to be 130 ka), one can assign "times" to the duration of the other soils' formation and, by inference, to the faulting events.

This approach was used by Machette (1978) to analyze faulting events at the Bernalillo County dump, which is west of Albuquerque in the north-central part of New Mexico. Here, a sequence of calcic soils (Bk and K horizons) have formed in response to faulting, and eolian and colluvial

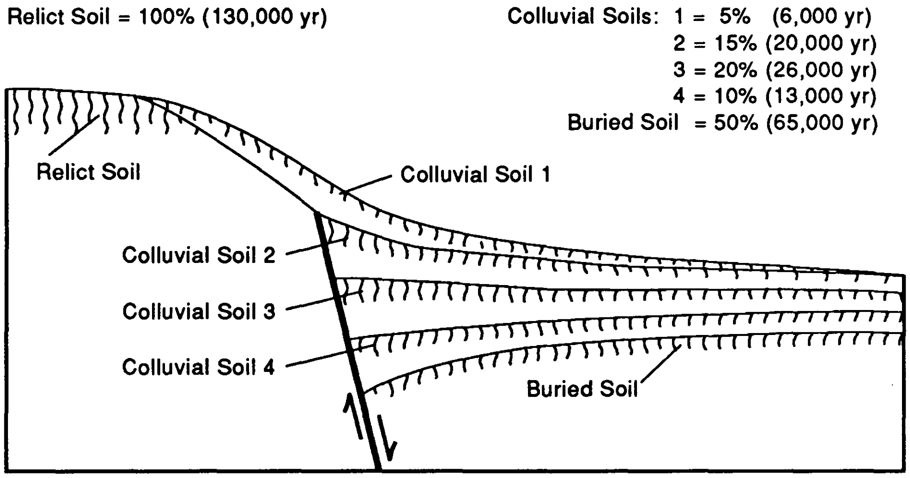


Figure 6-9. Schematic cross section of fault in surficial deposits. Sketch shows how estimates of times of faulting could be made from quantitative measures of soil development in relict and buried soils. Times shown in parentheses are lengths of soil formation. Model does not include time required for formation and stabilization of fault-scarp derived colluvium; other studies indicate that these processes may require several thousand years per faulting event.

deposition. Figures 6-10 through 6-12 show the stratigraphic relations and numerical calculations for four discrete faulting events that have occurred in the past 500,000 yr at this site. One basic assumption of his technique is that the rate of carbonate accumulation is linear through time, but the rate of accumulation is known to vary through time in relation to the supply of Ca^{2+} and the amount and distribution of rainfall (Machette, 1985a; McFadden and Tinsley, 1985). Thus, the estimated times of faulting could be in error if the incremental (20,000-100,000 yr) rates of accumulation differ from the long-term (500,000 yr) rate (Machette, 1985a). Nevertheless, this type of analysis allows one to evaluate, in general, the paleoseismic history of some normal faults. One could apply similar approaches to analyses of colluvium on hillslopes and to periodic eolian activity.

Nelson and Weisser (1983) used several measures of soil development (such as color indices, percent and total mass of clay and carbonate, and maximum percent of weathered clasts) to estimate the age of alluvial and glacial deposits in the Towanta Flat area of northeastern Utah. These deposits have been displaced, progressively through time, by the Uinta Basin fault. Using the soil-age estimates (Nelson and Weisser, 1983, table 1), they estimated a minimum time of 60 to >150 ka for the most recent faulting event on the Uinta Basin fault and an average recurrence interval of 50,000 to 100,000 yr for it during the middle Pleistocene.

Similar approaches, but dealing with latest Pleistocene and Holocene soils, have been used to study faulting in the back valleys of Utah (the extensional intermountain basin east of the Wasatch Range). In two of many studies by the Bureau of Reclamation, Nelson and VanArsdale (1986) and Sullivan and Nelson (1987) demonstrated recurrent late Quaternary movement on the normal faults in the Strawberry and Morgan Valleys. Such studies demonstrate the utility of soils and developmental indices as relative age indicators for tectonic events.

Soil A horizons also provide datable material for studies of fault chronology, both in a relative sense (Machette and Lund, 1987; Machette, 1988) and an absolute sense. Most often, radiocarbon age determinations are made from the organic carbon in tectonically buried A horizons, which may or may not be associated with charcoal or other burned woody material. However, in arid and semiarid climates, vegetation is sparse and charcoal is rare in either natural or artificial exposures. Machette and others (1987, 1991) used AMRT (apparent mean residence time) dates from A horizons along the Wasatch fault zone to limit the times of multiple Holocene faulting events (Fig. 6-13). They applied a correction to the AMRT dates to estimate the mean age of the soil's organic carbon at the time of burial, thereby obtaining more accurate estimates of the time of burial of the A horizons. In addition, silt from A horizons on colluvial wedges and from loess was sampled for experimental thermoluminescence dating (Forman and others, 1989, 1991). The combination of AMRT and conventional radiocarbon age determinations and TL age estimates from equivalent stratigraphic positions provided a valuable test of the TL method for soils, extended TL's usefulness as a dating tool for arid and semiarid environments, and demonstrated TL's ability to date faulting events.

Evidence of Structural Relations

The morphology and horizonation of soils can provide a tool for deciphering complex structural relations within and

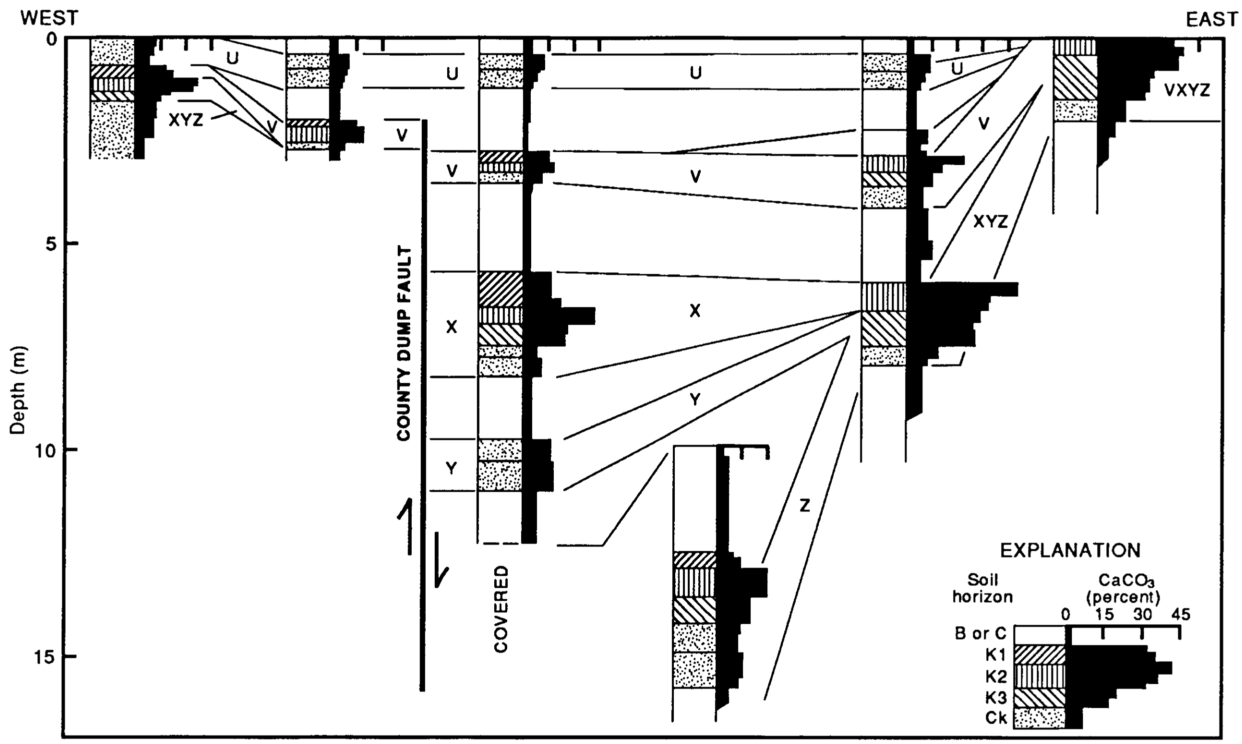


Figure 6-10. Cross section of relict and faulted buried soils at the Bernalillo County dump, New Mexico (from Machette, 1978). Section shows stratigraphic and structural position of soils that have been buried by colluvium and eolian sand as a result of normal slip on the fault. Small graphs show calcium carbonate content of soils *versus* soil horizonation.

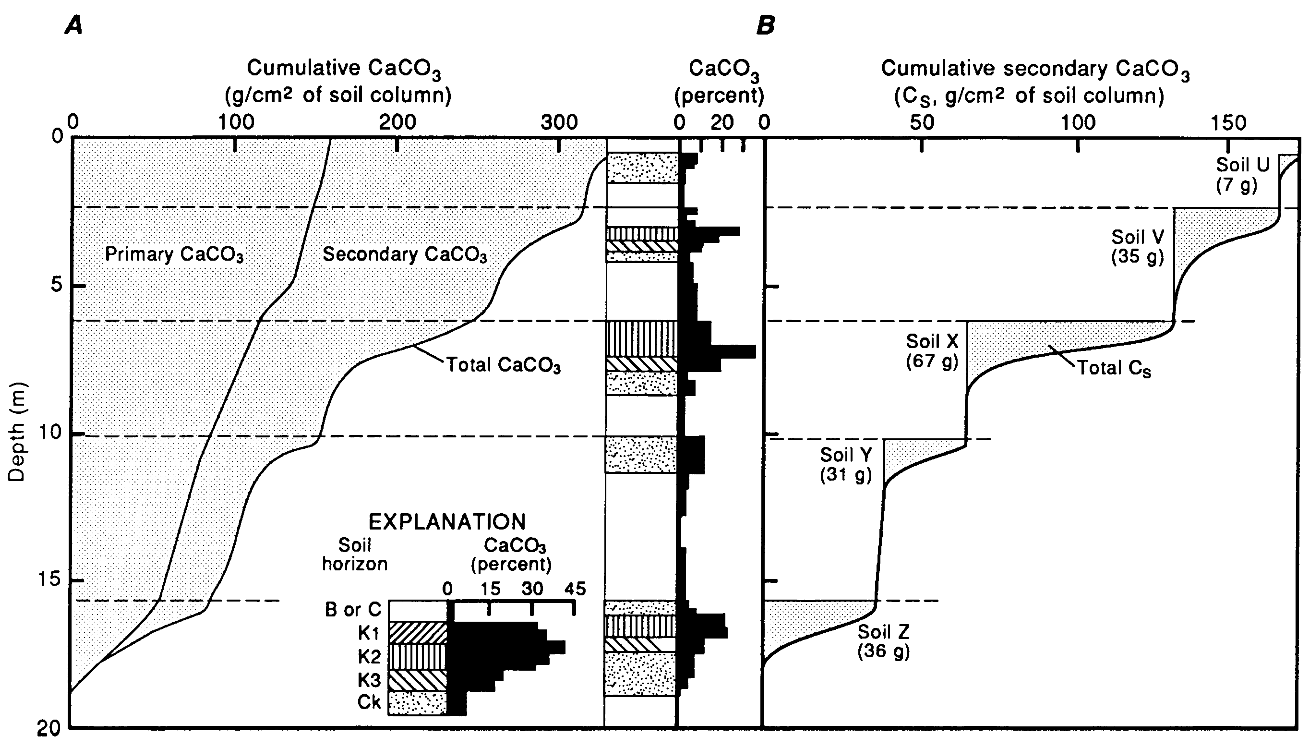
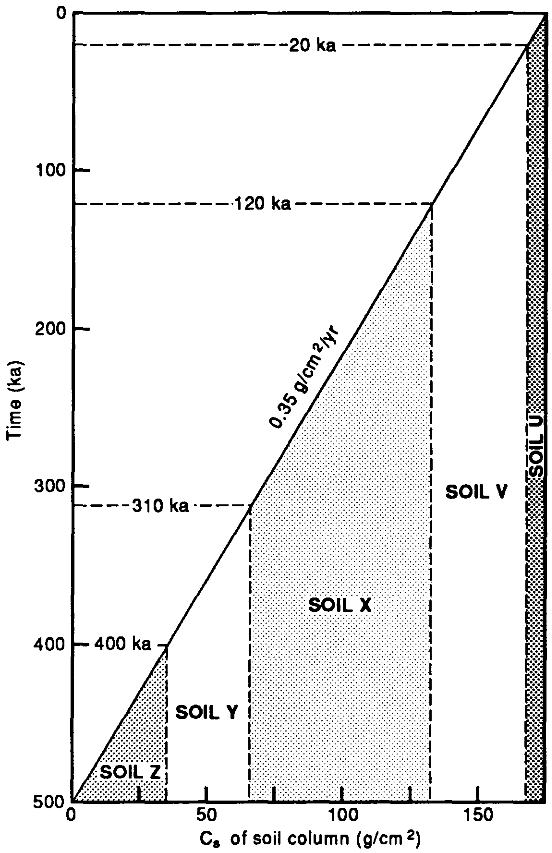


Figure 6-11. Calcium-carbonate data (g $CaCO_3/cm^2$) for soils exposed at the Bernalillo County dump, New Mexico (from Machette, 1978). *A.* Total, primary, and secondary calcium carbonate content of soils in composite section. *B.* Secondary calcium carbonate content of soils in composite section.



Fault Event	Time (ka)	Recurrence Interval (10^3 yr)	Displacement (m)
1	400	-----	>7.8
2	310	90	4.1
3	120	190	3.0
4	20	100	2.1

Figure 6-12. Timing of faulting events at the Bernalillo County dump as determined from secondary carbonate content and assumed constant rate of carbonate accumulation (from Machette, 1978).

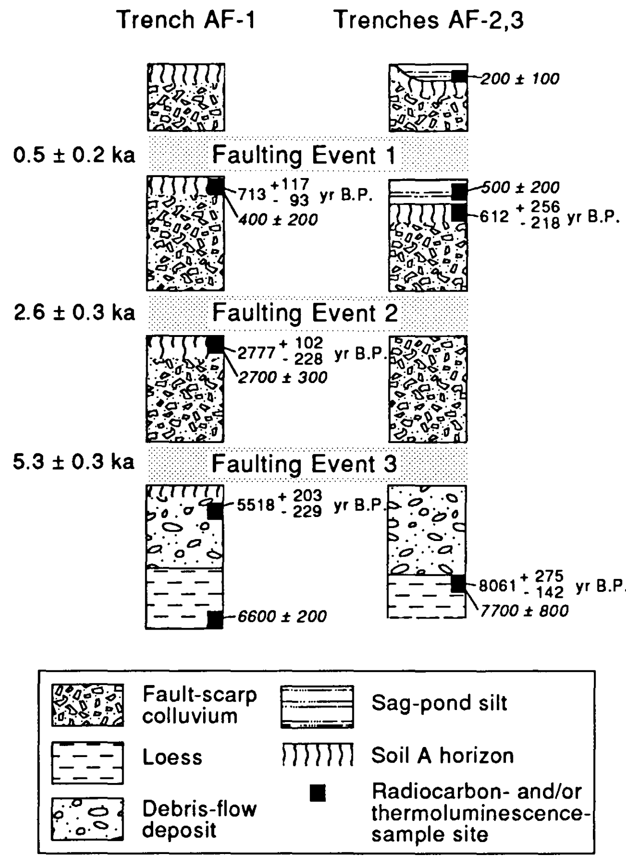


Figure 6-13. Possible controls on the timing of faulting events based on age determinations from soils. This example shows the results of dating of charcoal in and on A horizons, organic matter in A horizons, and TL age estimates from silt in bioturbated A horizons (from Forman and others, 1989). TL age estimates are shown in italics.

immediately adjacent to faults. For example, if the soil on the upthrown fault block has a well differentiated profile, then the orientation of disturbed blocks of this soils can be recognized in trenches cut across faults. In addition, the long-axis dimension of rotated soil blocks in colluvium can be used to calculate the minimum height of a fault's free face and, thus, estimate the amount of slip for the faulting event. For example, in Figure 6-14 the three soils on the upthrown fault block have a distinctive profile sequence consisting of Btkb3/Kb3/Btb4/Bkb4/Btb5 horizons (Machette, 1988). These horizons are part of a thin sequence of distal alluvial-fan deposits, each of which has a well developed soil. The Btkb3 and Kb3 horizons are part of an exhumed soil, whereas the Btb4/Bkb4/Btb5 horizon sequence is from the two oldest buried soils. The prominent rotated block (Btkb3/Kb3/Btb4/Bkb4) has fallen from the upthrown block as a result of undermining of the Bkb4 horizon in the previous free face of the scarp fault (3.6-4.0 m on horizontal axis, Fig. 6-14). The initial geometry of the fault scarp can be evaluated by rotating the blocks to their original positions and by restoring the volume of the coeval colluvium (light stipple pattern, Fig. 6-14) to the upthrown block, thereby revealing the probable scarp height and amount of tectonic displacement.

Estimates of Slip Rates from Soil/Alluvial Stratigraphy

Most reported slip rates for normal faults in the Basin and Range province are based on amount of displacement in either dated Holocene deposits (*i.e.*, Fig. 6-13) or in deposits that are correlated to dated deposits elsewhere in the region. An example of the latter is the slip rates for faulted deposits of the Bonneville lake cycle in northern Utah and eastern Nevada (Machette and others, 1987, 1991). Generally, slip rates are calculated for two well-known shorelines of this lake cycle—the Bonneville at 14.5 ka and the Provo at 14.2 ka—that are commonly displaced by the normal faults in the eastern Great Basin. However, slip rates for older datums

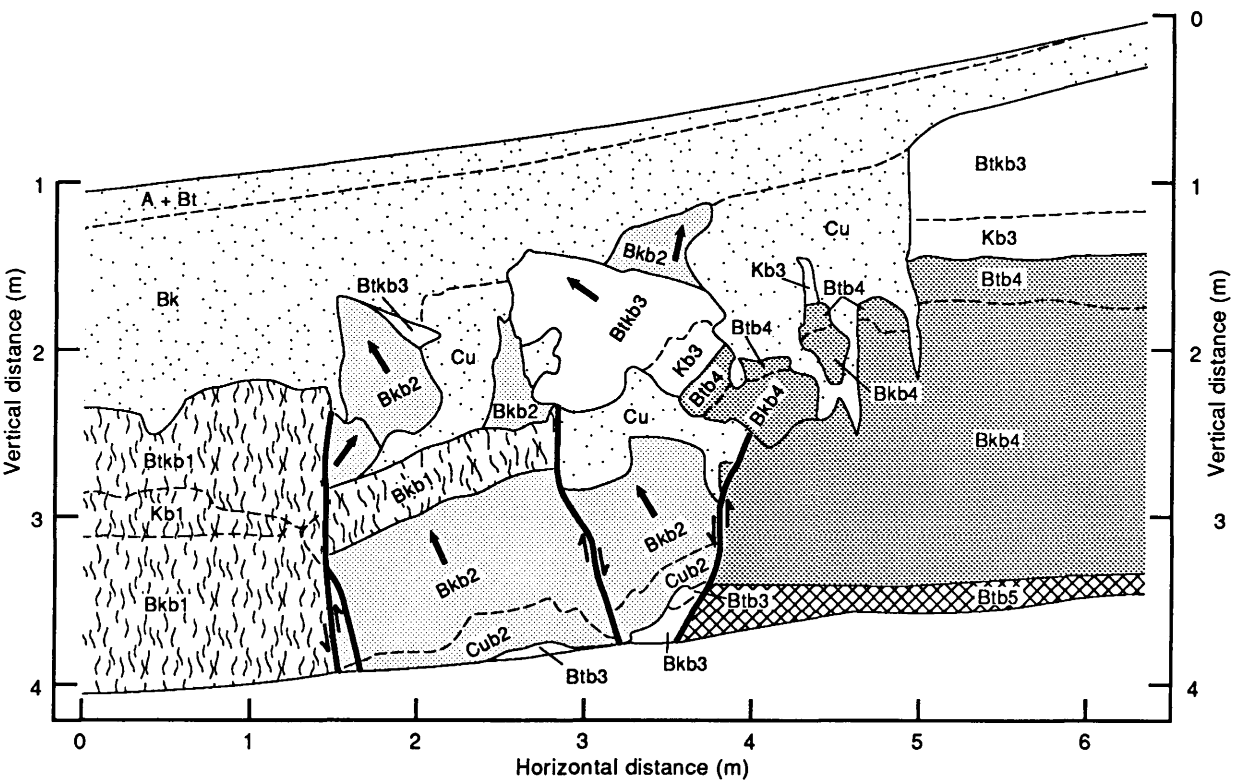


Figure 6-14. Map of trench across La Jencia fault, New Mexico (from Machette, 1988). This example illustrates the use of soil horizonation and stratigraphy to document catastrophic faulting events and the processes of fault scarp degradation.

(>20 ka) are rarely based on dated material, and often are inferred from the correlation of deposits with a climatic-time model that is tied to marine-oxygen-isotope records. As you can see, this is a rather tenuous chain of inferences.

Measures of soil development, either using PDI or total accumulation data, can be used to make reasonable age estimates for faulted deposits that are beyond the range of radiocarbon dating. For example, the Beaver fault (Machette, 1985b) cuts a sequence of flood-plain, terrace, and piedmont gravels along the eastern side of the Beaver basin (Machette, 1983). Repeated movement on the fault during the past 500,000 yr (Fig. 6-15) has yielded a 25 ± 5 m scarp on middle Pleistocene gravel (Qglc, Fig. 6-7), a 11 ± 2 m scarp on a late middle Pleistocene terrace (Qnc, Fig. 6-7), and a 1-3 m scarp on the latest Pleistocene alluvial flood plain (Qbv, Fig. 6-7). Age assignments for the soils on these units (Table 6-2) provide a maximum limit for the cumulative time of fault movement and, thus, yield minimum slip rates of 0.38 and 0.56 mm/yr for the late Pleistocene and middle Pleistocene, respectively (Fig. 6-15), and 0.50 mm/yr for the entire 500,000 yr. Although these estimates are subject to various errors, they provide a basis on which to evaluate the relative activity of faulting over broad regions and time scales not traditionally used in studies of paleoseismicity.

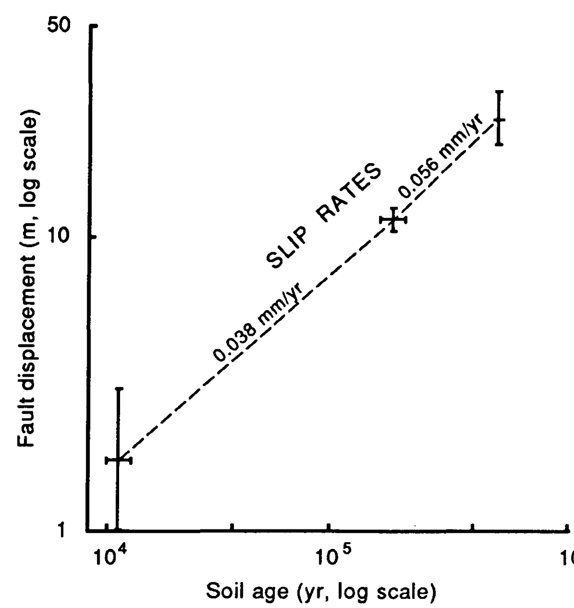


Figure 6-15. Example of estimated fault-slip rates based on amounts of offset recorded by scarps on different ages of alluvium in the Beaver basin. Age control (horizontal crossbars) comes from quantitative measures of soil development (see Table 6-2 and Fig. 6-8). Amounts of offset (vertical cross bars) are from field measures and fault-scarp profiles (Machette, unpubl. data, 1985).

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APPENDIX
DESCRIBING SOIL PROPERTIES

HORIZON NOMENCLATURE

It should be emphasized that horizon nomenclature depends not only on the properties of the particular horizon, but also on those of both the overlying and underlying horizons and the parent material (or the presumed parent material). Hence, it is not uncommon to change horizon designations in the soil description process, because the identification of surface horizons may change after more is learned of the adjacent horizons and the parent material.

There are certain rules to follow when more than one lower-case letter is used with the master horizon designation. In most cases, more than one lower-case letter can be used. These letters are always written first: a, e, i, h, r, s, t, v, and w. Further, none of the these letters is used in combination, except for Bhs and Crt. If more than one lower-case letter is used and the horizon is not buried, the following, if used, are commonly written last: c, f, g, m, and x. If used, b commonly is last, unless field or laboratory data indicate horizon features that formed after burial. For example, if carbonate accumulates in a Bt horizon after burial, the designation is Btbk. For B-horizon designation, t has precedence over w, s, and h, and the latter symbols are not used in combination with t. If other lower-case letters are used with t, the t comes first (*e.g.*, Btg).

SOIL PROPERTIES

The descriptive terminology for horizons developed by soil scientists is used to describe soils. An example of a soil-profile description is given in Table A-1. A field recording sheet that we have found useful is given in Table A-2; because describing soils can be quite messy, all one has to do is circle the correct property. In the text that follows, all of the abbreviations in Tables A-1 and A-2 are defined. Because these data are used to calculate the PDI, one person should collect all of the data for all of the soils of a particular project (see Chapter II).

Field site characteristics should be noted at the top of Table A-2. The configuration of the landform associated with the soil should be described, as concave parts of the the landscape can receive sediment or additional water from upslope. In contrast, convex parts of the landscape might be undergoing erosion. Ruhe (1975) has devised a classification useful for this purpose (Fig. A-1). The key properties of soils (Soil Survey Staff, 1975, 1981) are described in the following section.

Table A-1.—Field description of a soil formed at the summit (SU) position on a Bull Lake moraine, Idaho [Abbreviations are shown in Appendix. From Berry (1987, Table 2)]

Horizon	Depth (cm)	Horizon boundary	Munsell color (dry)	>2 mm fraction (%)	Soil texture
O	2-0	—	—	—	—
A	0-4	a,s	10*YR* 4.5/2	27	SL
AB	4-8	a,s	10*YR* 5/3	18	SL
Bw	8-26	a,s	10*YR* 6/3	53	SL
2Bt1	26-48	g,w	10*YR* 6/3	62	SL
2Bt2	48-84	g,w	10*YR* 6/4	64	SL
2Btj	84-125+	—	10-7.5*YR* 6/4	56	LS

Table A-1.—Continued
[Under parent materials, Eo/till indicates a mixed parent material of eolian material and till]

Horizon	Consistence		Structure	Clay films	Parent material
	Wet	Moist			
O	—	—	—	—	—
A	so,ps	lo	1f,gr	o	Eo/till
AB	so,ps	lo	1f,sbk	o	Eo/till
Bw	so,ps	lo	1f,sbk	o	Eo/till
2Bt1	ss,ps	fi	2m,sbk	2d br, 2d pf	Till
2Bt2	ss,ps	vfi	2m,abk	3d br, 2d pf	Till
2Btj	ss,po	fi	1m,sbk	1f br, 1f pf	Till

Table A-2.—Work sheet for recording soil properties in the field
[In the note column, one can record properties not universal to all soils. Courtesy of D. Jorgenson, 1989]

Soil Description: Location ______________________________

Site No. _____ Date _________ Time ___________ Vegetation ______________________________

Elevation __________ Slope ___________ Aspect _________ Geomorphic Surface ______________________________

Parent Material(s) ______________________________ Described by ______________________________

Depth (cm)	Horizon	Color moist / dry	Structure	Gravel %	Consistence Wet	Moist	Dry	Texture	pH	Clay films	Boundaries	notes
			m vf gr sg f pl 1 m pr 2 c cpr 3 vc abk sbk	0 50 <10 75 10 >75 25	so po ss ps s p vs vp	lo vfr fr fi vfi efi	lo so sh h vh eh	S SiCL LS SiL SL Si SCL SiC L C CL SC		v1 f pf 1 po 2 d br 3 co p cobr	a s c w g i d b	
			m vf gr sg f pl 1 m pr 2 c cpr 3 vc abk sbk	0 50 <10 75 10 >75 25	so po ss ps s p vs vp	lo vfr fr fi vfi efi	lo so sh h vh eh	S SiCL LS SiL SL Si SCL SiC L C CL SC		v1 f pf 1 po 2 d br 3 co p cobr	a s c w g i d b	
			m vf gr sg f pl 1 m pr 2 c cpr 3 vc abk sbk	0 50 <10 75 10 >75 25	so po ss ps s p vs vp	lo vfr fr fi vfi efi	lo so sh h vh eh	S SiCL LS SiL SL Si SCL SiC L C CL SC		v1 f pf 1 po 2 d br 3 co p cobr	a s c w g i d b	
			m vf gr sg f pl 1 m pr 2 c cpr 3 vc abk sbk	0 50 <10 75 10 >75 25	so po ss ps s p vs vp	lo vfr fr fi vfi efi	lo so sh h vh eh	S SiCL LS SiL SL Si SCL SiC L C CL SC		v1 f pf 1 po 2 d br 3 co p cobr	a s c w g i d b	
			m vf gr sg f pl 1 m pr 2 c cpr 3 vc abk sbk	0 50 <10 75 10 >75 25	so po ss ps s p vs vp	lo vfr fr fi vfi efi	lo so sh h vh eh	S SiCL LS SiL SL Si SCL SiC L C CL SC		v1 f pf 1 po 2 d br 3 co p cobr	a s c w g i d b	
			m vf gr sg f pl 1 m pr 2 c cpr 3 vc abk sbk	0 50 <10 75 10 >75 25	so po ss ps s p vs vp	lo vfr fr fi vfi efi	lo so sh h vh eh	S SiCL LS SiL SL Si SCL SiC L C CL SC		v1 f pf 1 po 2 d br 3 co p cobr	a s c w g i d b	
			m vf gr sg f pl 1 m pr 2 c cpr 3 vc abk sbk	0 50 <10 75 10 >75 25	so po ss ps s p vs vp	lo vfr fr fi vfi efi	lo so sh h vh eh	S SiCL LS SiL SL Si SCL SiC L C CL SC		v1 f pf 1 po 2 d br 3 co p cobr	a s c w g i d b	
			m vf gr sg f pl 1 m pr 2 c cpr 3 vc abk sbk	0 50 <10 75 10 >75 25	so po ss ps s p vs vp	lo vfr fr fi vfi efi	lo so sh h vh eh	S SiCL LS SiL SL Si SCL SiC L C CL SC		v1 f pf 1 po 2 d br 3 co p cobr	a s c w g i d b	
			m vf gr sg f pl 1 m pr 2 c cpr 3 vc abk sbk	0 50 <10 75 10 >75 25	so po ss ps s p vs vp	lo vfr fr fi vfi efi	lo so sh h vh eh	S SiCL LS SiL SL Si SCL SiC L C CL SC		v1 f pf 1 po 2 d br 3 co p cobr	a s c w g i d b	
			m vf gr sg f pl 1 m pr 2 c cpr 3 vc abk sbk	0 50 <10 75 10 >75 25	so po ss ps s p vs vp	lo vfr fr fi vfi efi	lo so sh h vh eh	S SiCL LS SiL SL Si SCL SiC L C CL SC		v1 f pf 1 po 2 d br 3 co p cobr	a s c w g i d b	

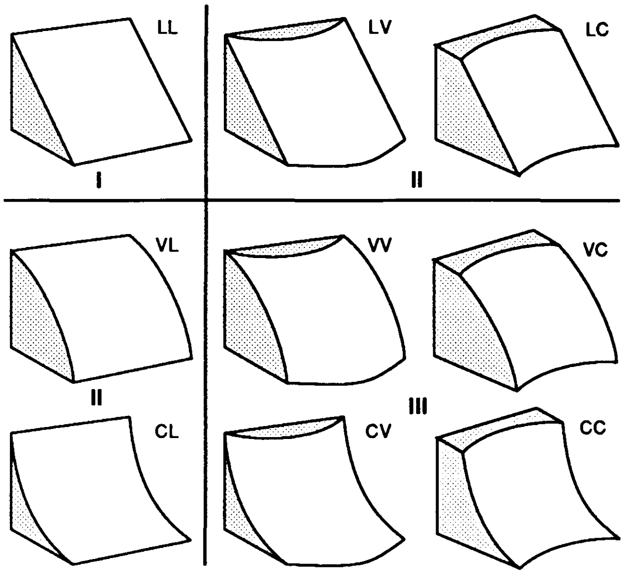


Figure A-1. Geometric forms of hillslopes. Slope length is down the form; slope width is across the form: L, linear; V, convex; and C, concave. The simplest form (I) if colinear (LL); group **III** forms, the most complex, are doubly curved; group **II** forms are linear in one dimension and curved in the other (from Ruhe, 1975).

Depth—The top of the uppermost mineral horizon (A or E) is taken as zero depth. The O-horizon thickness is measured up from that point (*i.e.*, 2-0 cm), and all other horizons down from that point (*i.e.*, 0-8 cm).

Color—List dominant color and size, and color variation of prominent mottles. For many studies, we prefer dry colors of sieved soil samples, all taken on the same day outdoors in direct sunlight. Use the Munsell Soil Color Charts (1954) or other suitable charts that use the Munsell color notation. List the moisture state when taken. If a color lies between two hue pages or color chips, this can be denoted, for example, as 7.5-10*YR* or 10*YR* 6/3.5, respectively. For mottled soils, record:

1. Where two or more colors occur in an intricate pattern but, if randomly mixed, indicate color as mottled 2.5*YR* 5/3, 5*Y* 4/4, *etc.*
2. Where one color is the continuous phase or matrix with mottles contained in it, designate this as the dominant color and specify the abundance, size, and contrast of the mottles as indicated below:

Abundance

f—few (mottles <2% of surface area).
c—common (mottles 2-20% of surface area).
m—many (mottles >20% of surface area).

Size

1—fine (<5 mm in diameter).
2—medium (5-15 mm in diameter).
3—large (>15 mm in diameter).

Scale, in mm

Contrast

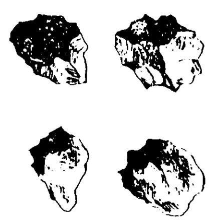
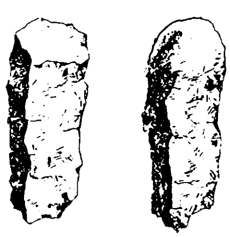
f—faint. Evident only on close examination. Faint mottles commonly have the same hue as the matrix color to which they are compared and differ by no more than 1 unit of chroma or 2 units of value. Some faint mottles of similar but low chroma and value differ by 2.5 units (one page) of hue.

d—distinct. Readily seen but contrast only moderately with the matrix color to which they are compared. Distinct mottles commonly have the same hue as the color to which they are compared but differ by 2 to 4 units of chroma and 3 to 4 units of value; or differ from the chroma to which they are compared by 2.5 units (one page) of hue but no more than 1 unit of chroma or 2 units of value.

p—prominent. Contrast strongly with the matrix color to which they are compared. Prominent mottles are commonly the most obvious color feature of the section described. Prominent mottles that have medium chroma and value commonly differ from the color to which they are compared by at least 5 units (two pages) of hue if chroma and value are the same, at least 4 units of value or chroma if the hue is the same, or at least 1 unit of chroma or 2 units of value if hue differs by 2.5 units (one page).

Color is a valuable aid, if used with caution, in qualitatively recognizing processes that are or have been operating in a soil. Indeed, color is the property that first catches one's attention with buried soils. Dark-brown to black colors in near-surface horizons reflect an accumulation of humified and (or) nonhumified organic matter. Dark colors may also result from the accumulation of MnO_2, but these usually have a bluish cast and commonly are not always close to the surface. Grayish colors (chromas) near 1 and (or) hues bluer than 10*Y* indicate reducing conditions (gleying), and color is due either to ferrous iron compounds or to the complete removal of free iron from the soil horizon. Yellow-brown to red colors result from the presence of iron oxides and hydroxides and are characteristic of B and C horizons. White or light-gray colors above the B horizon characterize an E horizon and suggest enough leaching by vertically or laterally moving water so that most of the grains are free of colloidal coatings of oxides and hydroxides of aluminum and iron. Similar colors below a Bw or Bt horizon are usually due to concentrations of $CaCO_3$. Soils that have been underwater or buried and beneath the water table may exhibit strange colors due to diagenesis.

Table A-3.—Types of soil structures

TYPE	SKETCH†	DESCRIPTION	PROBABLE ORIGIN	ASSOCIATED WITH THESE SOIL HORIZONS
Granular (g)		Spheroidally shaped aggregates with faces that do not accommodate adjoining ped faces.	Colloids, mainly organic, bind the particles together; clay and Fe and Al hydroxides may be responsible for some binding, and flocculating capacity of some ions, such as Ca^{2+}, may be helpful; periodic dehydration helps form more stable aggregates.	A
Angular blocky (abk) Subangular blocky (sbk)		Approximately equidimensional blocks with planar faces that are accommodated to adjoining ped faces; face intersections are sharp with angular blocky, rounded with subangular blocky.	Many faces may be intersecting shear planes developed during swelling and shrinkage that accompany changes in soil moisture.	Bt
Prismatic (pr, left) Columnar (cpr, right)		Particles are arranged about a vertical line, and ped is bounded by planar, vertical faces that accommodate adjoining faces; prismatic has a flat top, and columnar a rounded top.	Faces develop as a result of tension during times of dehydration; rounded column tops may be due to some combination of erosion by percolating water and greater amounts of upward swelling of column centers upon wetting; columnar commonly associated with the high exchangeable Na^{+} (Bn horizon).	Bt, Bn
Platy (pl)		Particles are arranged about a horizontal plane.	May be related to particle size orientation inherited from parent material or induced by freeze-thaw processes.	E, or those with fragipan
			May be related to layering in cementing material, induced during its precipitation (carbonate, silica, Fe hydroxides).	Km, Bqm, Bsm

† From Soil Survey Staff (1975)

Structure—Describe type, grade, structure size.

Type of structure: Use Table A-3 to define the type of soil structure.

Grade

- m—massive. Enough aggregation to maintain a vertical face but no formation of structure type.
- sg—single grain. No aggregation.
- 1—weak. Peds barely observable in place, and when disturbed, few entire peds are observed; much of the material is unaggregated.
- 2—moderate. Peds easily observable but not distinct in place, and when disturbed, many entire peds.
- 3—strong. Peds are distinctly visible in place, and when disturbed, nearly the entire mass consists of entire peds.

Size: Size differs with the kind of structure as shown in Table A-4.

Gravel content—Estimate volume percent occupied by gravel (>2 mm). Weight percent can be determined in the field with a screen (we use 3-mm door screen) and a hand-held portable scale. Be watchful for shape and lithologic changes during the screening process, as they may indicate parent materials of more than one origin.

Consistence—This is a measure of the adherence of the soil particles to the fingers, the cohesion of soil particles to one another, and the resistance of the soil mass to deformation. Because this property varies with moisture content, it is taken when the soil is dry, moist, and wet. The wet consistence (natural or artificial wetness) is useful in determining texture classes in the field.

Dry consistence

- lo—loose. Noncoherent.
- so—weakly coherent. Easily crushes to powder or single grain.
- sh—slightly hard. Easily broken between thumb and forefinger.
- h—hard. Can be broken in the hands without difficulty; difficult to break between thumb and forefinger.
- vh—very hard. Can be broken in hands with difficulty.
- eh—extremely hard. Cannot be broken in hands.

Moist consistence

- lo—loose. Noncoherent.
- vfr—very friable. Crushes under gentle pressure.
- fr—friable. Crushes easily under gentle to moderate pressure between thumb and forefinger.
- fi—firm. Crushes under moderate pressure between thumb and forefinger but resistance is distinctly noticeable.
- vfi—very firm. Crushes under strong pressure, barely crushable between thumb and forefinger.
- efi—extremely firm. Crushes under very strong pressure, cannot be crushed between thumb and forefinger.

Wet consistence

<u>Stickiness</u> is measured by pressing the wet soil between the thumb and forefinger and noting its adherence.

- so—nonsticky. Practically no adherence when pressure released.
- ss—slightly sticky. After pressure, soil adheres to both thumb and finger but comes off one rather cleanly. Does not appreciably stretch.
- s—sticky. After pressure soil adheres to both thumb and finger and tends to stretch somewhat before pulling apart from either digit.
- vs—very sticky. After pressure, soil adheres strongly to both digits and is markedly stretched when they are separated.

<u>Plasticity</u> is measured by rolling the wet soil between the thumb and finger and observing whether or not a wire or thin rod can be formed.

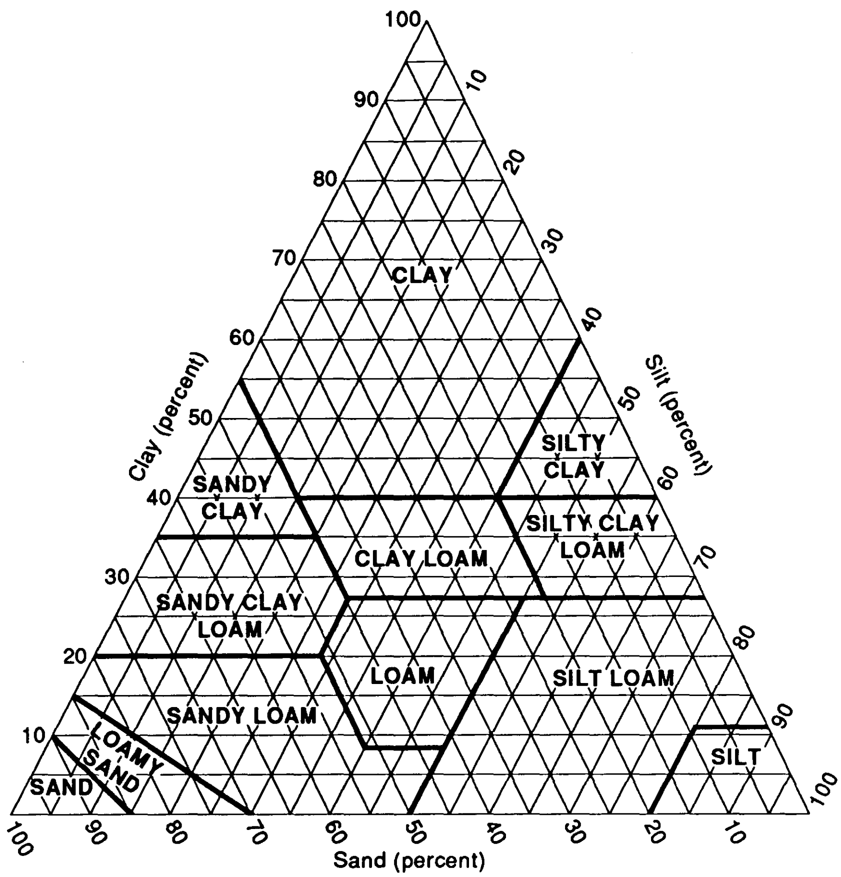
- po—nonplastic. No thread can be formed.
- ps—slightly plastic. Thread can be formed and mass is deformed by very slight force.
- p—plastic. Thread can be formed and mass is deformed by slight force.
- vp—very plastic. Thread can be formed and mass is deformed by moderate or strong force.

Table A-4.—Classes of soil structures

0 20 40

Scale, in mm

SIZE CLASS	DIAMETER OF GRANULES (mm)	THICKNESS OF PLATES (mm)	DIAMETER OF BLOCKS (mm)	DIAMETER OF PRISMS (mm)
vf—very fine	<1	<1	<5	<10
f—fine	1-2	1-2	5-10	10-20
m—medium	2-5	2-5	10-20	20-50
c—coarse	5-10	5-10	20-50	50-100
vc—very coarse	>10	>10	>50	>100



TEXTURAL ABBREVIATIONS:			
C	Clay	SCL	Sandy Clay Loam
CL	Clay Loam	SL	Sandy Loam
L	Loam	Si	Silt
LS	Loamy Sand	SiC	Silty Clay
S	Sand	SiCL	Silty Clay Loam
SC	Sandy Clay	SiL	Silt Loam

MODIFIER ABBREVIATIONS:	
vf	very fine
f	fine
co	coarse
vco	very coarse
g	gravelly

Figure A-2. Textural names and abbreviations of names *versus* sand-silt-clay contents.

Texture—Use established names from the textural triangle (Fig A-2). Determine the textural class of the less than 2-mm fraction by noting the grittiness and wet consistence in a rating chart (Table A-5). Broad guidelines are given in the rating chart, but for more accuracy, one should calibrate the tactile response by texturing samples with known particle-size distribution.

pH—Record specific value, using field kit.

Clay films—These are described by recording their amount, distinctness, and locations. Study with a hand lens in the field, or with a microscope in the laboratory.

Amount

v1—very few. Occupies less than 5% of the total area of the kind of surface described.

1—few. Occupies 5-25% of the total area of the kind of surface described.

2—common. Occupies 25-50% of the total area of the kind of surface described.

3—many. Occupies more than 50% of the total area of the kind of surface described.

The same classes are used to describe the amount of bridges connecting particles of structureless soil bodies. The amount is judged on the basis of the percentage of particles of the size designated that are joined to adjacent particles of similar size by bridges at contact points.

Distinctness: Distinctness refers to the ease and degree of certainty with which a surface feature can be identified. Distinctness is related to thickness, color contrast with the adjacent material, and other properties but is not itself a measure of any one of them. Some thick films, for example, are faint, whereas some thin ones are prominent. The distinctness of some surface features changes markedly as

Table A-5.—Field criteria useful for determining major textural classes
[Modified from Foss and others (1975)]

Soil textural classes	Feel (moist)	Ability to: Form stable ball	Ribbon out	"Soils Hands"	Plasticity	Stickiness	Consistence Moist	Dry
Sand	Very gritty	No	No	No	No	No	Loose	Loose
Loamy sand	Very gritty	No	No	Yes (slight)	No	No	Loose	Loose
Sandy loam	Gritty	Yes (easily deformed)	Yes (dull surface; poorly formed)	Yes	No	No	Very friable	Weakly coherent
Loam	Gritty	Yes	Yes (dull surface; poorly formed)	Yes	Yes (slight)	Yes (slight to sticky)	Friable	Weakly coherent
Silt loam	Velvety	Yes	Yes (dull surface; poorly formed)	Yes	Yes (slight to plastic)	Yes (slight to sticky)	Friable	Weakly coherent
Silty clay loam	Velvety and sticky	Yes (very stable)	Yes (shiny surface; well formed)	Yes	Yes (plastic)	Yes (sticky)	Friable to firm	Slightly hard
Clay loam	Gritty and sticky	Yes (very stable)	Yes (shiny surface; well formed)	Yes	Yes (plastic)	Yes (sticky)	Firm	Slightly hard to hard
Sandy clay loam	Very gritty and sticky	Yes (very stable)	Yes (shiny surface; well formed)	Yes	Yes (plastic)	Yes (sticky)	Friable to firm	Slightly hard to hard
Silty clay	Extremely sticky and very smooth	Yes (very resistant to molding)	Yes (very shiny surface; very well formed)	Yes	Yes (very)	Yes (very)	Firm to very firm	Hard to very hard
Clay	Extremely sticky	Yes (very resistant to molding)	Yes (very shiny surface; very well formed)	Yes	Yes (very)	Yes (very)	Firm to very firm	Hard to very hard

Note: All criteria, except for the right two columns, are for soil in the wet state, but not so wet that it flows off one's hand. These criteria can be used as a "general guide" to determine field textures, but local variations in organic matter, clay mineralogy, carbonate, parent material, and other factors would require adjustments before their use in a given area.

the amount of moisture changes; therefore, the soil-water state is specified. These distinctness classes are used.

f—faint. Evident only on close examination with 10X magnification and cannot be identified positively in all places without greater magnification. The contrast with the adjacent material in color, texture, and other properties is small.

d—distinct. Can be detected without magnification, although magnification or tests may be needed for positive identification. The feature contrasts enough with the adjacent material that a difference in color, texture, or other properties is evident.

p—prominent. Conspicuous without magnification when compared with a surface broken through the soil. Color, texture, or some other property or combination of properties contrasts sharply with properties of the adjacent material, or the feature is thick enough to be conspicuous.

Location of clay films: Oriented clay is present as films on peds, inside of pores, or as bridges between grains and coats on grains.

pf—clay films occur on ped faces. Where the structure grade is weak or the soil is structureless, ped faces are indistinct or absent. It is probable that only when the structure grade is moderate or strong are the clay films on ped faces are discernible.

po—clay films line tubular or interstitial pores.

br—oriented clay occurs as bridges holding mineral grains together. (This is probably an initial step that occurs before clay films coat grains and is best observed in coarse-textured soils.)

co—colloid coats mineral grains.

cobr—coats and bridges are present (probably more common than coats or bridges alone).

In describing clay films, care must be exercised not to confuse pressure faces with clay films. Pressure faces may arise because of slickensides (caused by soil slip), swelling that pushes structural aggregates together and makes their sides look smooth (in places reflective) and striated, and enlargement of roots in tubular pores.

Examples of clay-film descriptions

3d po—many distinct clay films in pores.

2f pf & po—common faint clay films on peds and in pores.

3p pf, 2f po—many prominent clay films on ped faces, common faint clay films in pores.

It is important to record clay films because their presence is strong evidence for pedogenically illuviated clay. However, be warned that in places clay films can be original depositional (parent-material) features. Waters charged with fine sediment that infiltrate a flood plain can produce clay films at depth (Walker and others, 1978), as can similar waters infiltrating till at the base of a glacier. If these latter parent-material films are present below the main soil-forming zone, their color will be closer to that of the parent material than to that of the soil.

Horizon boundaries—Describe the lower boundary of each horizon, indicating distinctness and general topography.

Distinctness

a—abrupt. Transition is less than 2 cm.
c—clear. Transition is 2-5 cm thick.
g—gradual. Transition is 5-15 cm thick.
d—diffuse. Transition is more than 15 cm thick.

Topography: The modifiers sl (slightly) and v (very) may be used in combination with the following abbreviations.

s—smooth. Boundary is parallel to surface of the soil.
w—wavy. If pockets are wider than their depth.
i—irregular. If irregular pockets are deeper than their width.
b—broken. If parts of the horizon are unconnected with other parts.

Stages of carbonate morphology—Describe the stage of morphology with respect to the texture of the parent materials as described in Tables A-6 and A-7. In some places, there may not be stage II morphology in a sequence of nongravelly soils; rather, filaments of stage I become so common that the horizon meets the approximate percent requirements for stage II. Holliday (1982) suggests that these latter occurrences be termed IIf to point out their filamentous morphology.

We want to inject a word of caution on the recognition of carbonate morphological stages. In places, carbonate can be deposited on vertical faces by laterally seeping waters and thereby mask the pedogenic carbonate morphology (Lattman, 1973). In addition, Machette and R.E. Anderson (USGS) have observed strong lateral (on contour) variations in carbonate morphology and accumulation along natural arroyos in arid parts of the eastern Great Basin. Hence, to study the morphology of pedogenic carbonate and avoid surficial cementation, one may have to dig back a meter or more.

Table A-6.—Stages of carbonate morphology in gravelly and nongravelly parent materials
[Modified from Gile and others (1966); Bachman and Machette (1977); R.R. Shroba (written commun., 1982); and Machette (1985a)]

STAGE	GRAVELLY PARENT MATERIAL	NONGRAVELLY PARENT MATERIAL
I	Thin discontinuous clast coatings; some filaments; matrix can be calcareous next to stones; about 4% $CaCO_3$.	Few filaments or coatings on sand grains; <10% $CaCO_3$.
I+	Many or all clast coatings are thin and continuous.	Filaments are common.
II	Continuous clast coatings; local cementation of few to several clasts; matrix is loose and calcareous enough to give somewhat whitened appearance.	Few to common nodules; matrix between nodules is slightly whitened by carbonate (15-50% by area), and the latter occurs in veinlets and as filaments; some matrix can be noncalcareous; about 10-15% $CaCO_3$.
II+	Same as stage II, except carbonate in matrix is more pervasive.	Common nodules; 50-90% of matrix is whitened; about 15% $CaCO_3$.
Continuity of fabric high in carbonate		
III	Horizon has 50-90% of grains coated with carbonate, forming an essentially continuous medium; color mostly white; carbonate-rich layers more common in upper part; about 20-25% $CaCO_3$.	Many nodules, and carbonate coats so many grains that over 90% of horizon is white; carbonate-rich layers more common in upper part; about 20% $CaCO_3$.
III+	Most clasts have thick carbonate coats; matrix particles continuously coated with carbonate or pores plugged by carbonate; cementation more or less continuous; > 40% $CaCO_3$.	Most grains coated with carbonate; most pores plugged; >40% $CaCO_3$.
Partly or entirely cemented (irrespective of parent material)		
IV	Upper part to K horizon is nearly pure cemented carbonate (75-90% $CaCO_3$) and has a weak platy structure due to the weakly expressed laminar depositional layers of carbonate; the rest of the horizon is plugged with carbonate (50-75% $CaCO_3$).	
V	Laminar layer and platy structure are strongly expressed; incipient brecciation and pisolith (thin, multiple layers of carbonate surrounding particles) formation.	
VI	Brecciation and recementation (multiple generations), as well as pisoliths, are common.	

Salts: For salts more soluble than carbonates, use the carbonate morphology scheme.

Cementation—Refers to brittle, hard consistence caused by some cementing agent, such as silica or $CaCO_3$ which, unlike clay, does not deform under pressure.

cw—weakly cemented. Mass is brittle and hard, but can be broken in hands.

cs—strongly cemented. Mass is brittle, cannot be broken in hands, easily broken with hammer.

ci—indurated. Very strongly cemented, brittle, does not soften under prolonged wetting; breaks only with a sharp blow with a hammer, rings to hammer blow.

Cementation may be continuous or discontinuous and this feature should be described.

Silt caps—In places, one can detect translocation of silt and clay by the presence of these materials on the tops of clasts in the soil (Table A-8). The caps become thicker and cover a greater proportion of the clast surface with time (stages 1-4), form bridges between clasts (stage 5), and eventually encapsulate all clasts (stage 6). Stages 5 and 6 characterize the B horizon. Some silt caps can be parent-material features; these will have colors closer to those of the parent material than to those of the soil.

Table A-7.—Further subdivision of carbonate stages I and II in gravelly material on the basis of coating thickness and coverage
[Taken from Forman and Miller (1984)]

STAGE OF CARBONATE DEVELOPMENT	CHARACTERISTICS		
	Bottom coverage (percent)	Thickness (mm)	Remarks
IA	5-30	<0.5	Thin coats.
IB	30-70	≤0.5	
IC	70-100	0.5-1.0	
IIA	90-100	1-2	Thick coat; 0.5-1.0 mm pendants.
IIB	100	2-5	Thick plates; 1-3 mm pendants.

Table A-8.—Stages of silt-cap development and their characteristics
[Taken from Forman and Miller (1984)]

STAGE OF SILT CAP DEVELOPMENT	CHARACTERISTICS		
	Thickness (mm)	Coverage (percent)	Remarks
1 (cap)	<1	5-50	
2 (cap)	1-4	50-90	
3 (cap)	4-7	75-90	
4 (cap)	5-10	90-100	
5 (cap and bridge)	10-20	100	Caps interconnected to form bridges.
6 (encapsulate)	10-30	100	Clast completely encapsulated by infiltrated silt.