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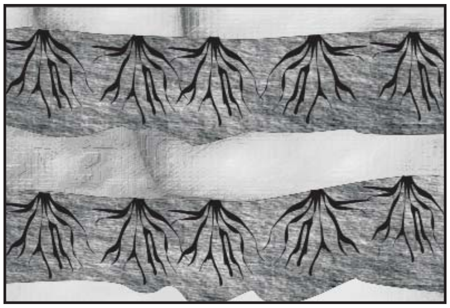
International Year of Planet Earth
Année Internationale de la Planète Terre
2008 - 2009

SERIES



International Year of Planet Earth 4. Utilizing Paleosols in Quaternary Climate Change Studies

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SUMMARY
Paleosols are an important tool in interpreting Quaternary stratigraphic sequences. Buried, exhumed, and relict soils delineate ancient surfaces that may have undergone weathering processes for long periods. Soils, including many buried soils, sometimes manifest complex histories characterized by polygenetic or superimposed profiles. Identification of the type of paleosol, or at least a diagnostic horizon, can provide valuable insight into past climates, which in turn assists in determining past vegetation. Interpreting a paleosol may be hindered by eroded or partially preserved horizons, complex climatic and environmental histories, and in the case of buried paleosols, alteration of one or more soil horizons by infiltration of material from overlying units. Soil thickness allows the minimum time of the weathering interval to be estimated and various methods are available for estimating the age of a paleosol. Probably the most accurate dating technique involves bracketing the age of the paleosol using datable material in sedimentary units above and below. Buried soils are found throughout the world, and are common in eolian deposits, such as loess and sand dunes, and alluvial deposits. In Canada, and other glaciated regions, they are interbedded with glacial deposits and have been most useful in determining periods of sedimentary nondeposition, and interpreting interglacial climates. The loess–paleosol sequences of the Chinese Loess Plateau, where over 37 major climatic cycles have been identified in the past 2.6 Ma, are presented as an outstanding example of the usefulness of buried soils. The Chinese paleosols are similar to Canadian chernozems and Luvisols, indicating steppe–forest to dense forest or steppe–forest vegetation. The cycles are best explained by Milankovitch forcing.

RÉSUMÉ
Les paléosols constituent un outil important dans l'interprétation des séquences stratigraphiques quaternaires. Les sols enfouis, exhumés et reliques constituent des surfaces anciennes qui peuvent avoir été soumises à l'altération pendant de longues périodes. Les sols, notamment les sols enfouis, révèlent parfois des histoires complexes caractérisées par des profils polygéniques ou superposés. L'identification du type de paléosol, ou d'un horizon diagnostique du moins, peut nous dire beaucoup sur les climats anciens, ce qui nous aide ensuite à connaître la végétation ancienne. L'interprétation des paléosols peut être compliquée du fait d'horizons érodés ou partiellement préservés, de la complexité des histoires climatiques et environnementales, et dans le cas de sols enfouis, par l'altération d'un ou de plusieurs horizons par infiltration provenant des couches sus-jacentes. L'épaisseur des sols permet d'estimer la durée d'altération minimum alors que des méthodes variées permettent d'en estimer l'âge. La technique de datation la plus juste probablement, consiste à délimiter une fourchette d'âge d'un paléosol à partir de matériel approprié des unités sédimentaires sus-jacents et sous-jacents. Il existe des sols enfouis partout dans le monde, et ils sont fréquents dans les dépôts éoliens, tel ceux des lœss et des dunes de sables, et les dépôts alluviaux. Au Canada et dans d'autres régions jadis glaciaires, les paléosols sont intercalés dans des dépôts glaciaires ont été très utiles pour déterminer les périodes d'absence de dépôts, et pour interpréter les climats interglaciaires. Les séquences lœss-paléosol du Plateau de lœss de Chine, où 37 cycles climatiques majeurs ont été circonscrits au cours des derniers 2,6 Ma, sont décrits à titre d'exemple remarquable de l'utilité des sols enfouis. Les paléosols de Chine sont semblables aux tchernozioms et luvisols canadiens, et représentent une végétation allant de la forêt de steppe à la forêt dense. Les cycles s'expliquent le mieux par l'hypothèse du forçage de Milankovitch.

INTRODUCTION
The use of paleosols has long been important in interpreting Quaternary history. When soils, buried or at the modern surface, are considered to be older than the soil that is developing at

the present time, they can be classified as paleosols. There are three types of paleosols:

i) Relict soils, which have remained on the current land surface since the time of initial formation but do not necessarily reflect the processes forming today's soil;
ii) Buried soils, which formed on ancient land surfaces that were subsequently buried by a younger deposit; and
iii) Exhumed soils, which were initially buried but were later exposed by erosion.

All of these types of paleosols are useful in climate-change studies, but buried soils have been most important in deriving climate information for a great number of Quaternary time intervals because they have remained protected from most alteration since their formation. The objective here is not to discuss all aspects of paleosols, but rather to present some of the author's experiences with using paleosols to reconstruct Quaternary climate change. There have been many investigations on pre-Quaternary paleosols in rocks as old as Precambrian (see Martini and Chesworth 1992). This paper will discuss what soils are, where paleosols are commonly found, how climate can be interpreted from paleosols, and problems of dating paleosols. A case study is presented in which paleosols have been particularly useful in climate-change investigations.

WHAT ARE SOILS?

Soils are commonly defined as dynamic bodies having properties derived from the combined effects of climatic and biotic activities, as modified by topography, acting on parent materials (rocks and sediments) over periods of time. They form an interface, or a skin, between the lithosphere, atmosphere, hydrosphere and biosphere (Brady and Weil 2002). The sun supplies the energy for the physical and chemical weathering that produces a soil. The complexity of nature ensures that soil characteristics will vary from place to place, and because of the changing nature of the soil-forming factors, no two soil profiles are exactly the same (Table 1; Fig.1).

A soil consists of a matrix of inorganic and organic solid particles in association with interconnected voids containing varying amounts of water and gases. Common defining characteristics of soils are colour, grain size, mineral content, diagenetic structure, organic matter content, bulk density, and soil moisture. The arrangement of these characteristics into horizons that form a soil profile is caused by varying degrees of weathering intensity.

In Canada there are four commonly used master soil horizons, designated O, A, B, and C (Canadian System of Soil Classification 1978). An O horizon is composed of organic matter that lies above the mineral soil. The A horizon is the top-most mineral horizon, containing partially decomposed

Table 1. Ranking of the importance of various processes in the genesis of soils of the Soil Orders. Explanation of letters in ranking: o - none to slight; l - little; m - moderate; h - high. Modified from Smeck et al. (1983).

Process	Regsol	Brunisol	Organic	Solonetz	Chernozem	Podzol	Luvisol
Physical mixing	l	l	o	l	m	o	o
Mineral weathering	l	l	o	l	m	m	m
Formation secondary minerals	o	l	o	m	l	m	m
Leaching	o	l	o	o	l	m	m
Eluvial - illuvial	o	l	o	h	m	h	h
Organic matter accumulation	o	m	h	o	m	m	m

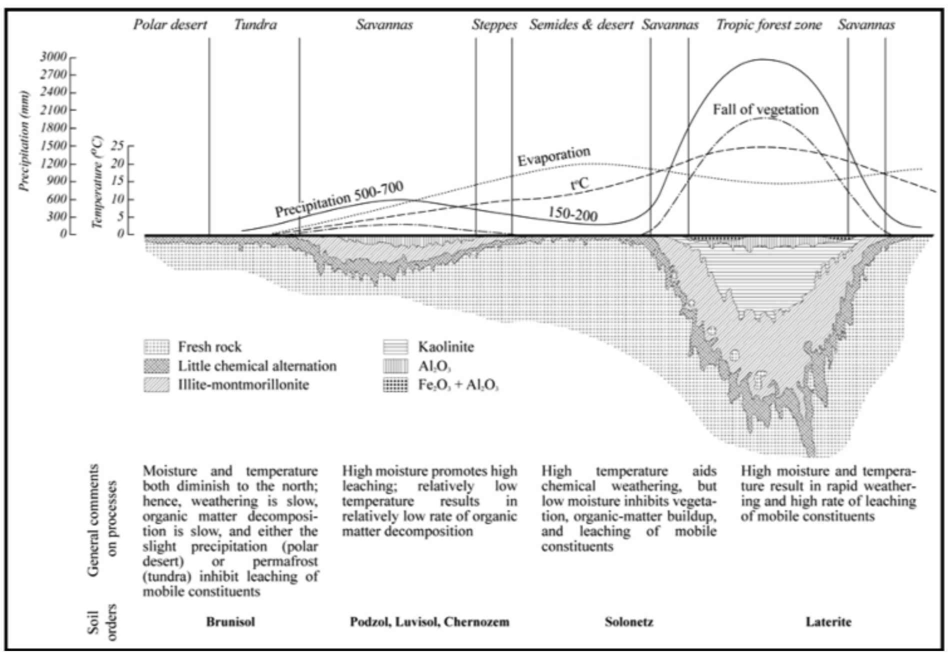

Figure 1. Diagram of relative depth of weathering and weathering products as they relate to some environmental factors in a transect from the equator to the north polar region. Modified from Strakov (1967).

mineral matter, but also usually contains abundant organic matter in its upper part. The B horizon has undergone extensive changes during soil genesis so that the original parent material is no longer discernible; clay minerals have commonly accumulated by being translocated from above, and new minerals may have formed in place via chemical and physical reactions. The C horizon is the least altered part of the soil and commonly resembles the original parent material, although it may contain secondary mineral accumulations. Collectively, the soil horizons that have undergone the same soil forming conditions are referred to as the solum (Fig. 2). Each of these master horizons can be subdivided with a letter or number that designates the presence of a distinctive feature, which aids in further identifying the soil (see section on Mineral Horizons, page 69).

In the Canadian System of Soil Classification (1978), there are nine soil Orders (Table 1):

1) Brunisolic, commonly found in moist forest regions of eastern and western Canada;
2) Chernozemic, typically occurring in the cool, subarid to subhumid climates of central Canada;
3) Crysolic, found in permafrost regions of northern Canada;
4) Gleysolic, developed in poorly drained areas with fluctuating water tables;
5) Luvisolic, typically found under forests or forest margins, in subhumid to humid, mild to very cold climates such as the central to northern Interior Plains;
6) Organic, composed largely of water-saturated organic materials such as peat, and found in poorly drained areas;
7) Podzolic, developed in acid terrain associated with forest and heath vegetation, in cool or very cold, humid climates such as the mountains of western Canada;
8) Regosolic, or partially developed soils; and
9) Solonetzic, found on saline parent materials in some areas of the semiarid to subhumid Interior Plains.

These Orders are further subdivided into more specific Great Groups and

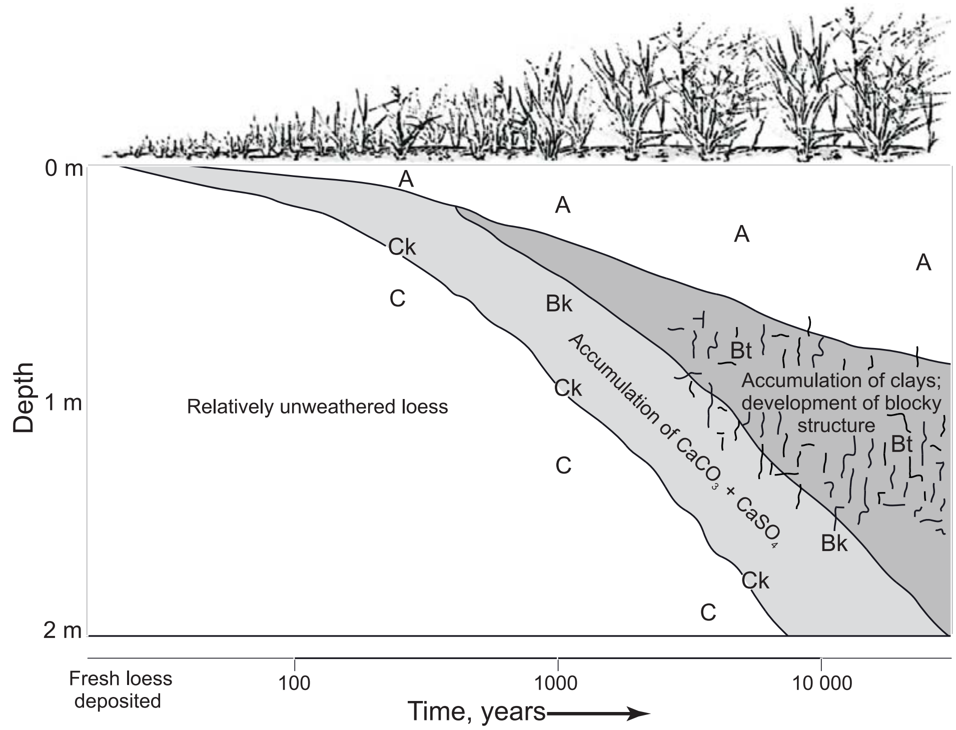

Figure 2. Variation through time of soil characteristics developed in calcareous loess in a warm subhumid climate supporting prairie grass vegetation. Initially, rainwater that is charged with organic acids dissolves carbonates and moves them down to a zone of carbonate accumulation (Bk or Ck horizon). Organic matter is incorporated and helps develop an A horizon. Over time the zone of carbonate accumulation moves deeper, the A horizon thickens and a noncalcareous B horizon develops. As the B horizon continues to develop (evolving to a Bt horizon), color changes occur, structure develops, and silicate clay accumulates both by weathering of primary minerals, and by movement by solutions from above. The mature chernozemic soil is on the right side of the diagram. Similar scenarios could be presented for other soil orders. Modified after Brady and Weil (2002).

Subgroups. For more information on soil characteristics, see Birkeland (1999), Brady and Weil (2002), Buol et al. (1997), and Fanning and Fanning (1989).

It can be difficult to determine how long a soil has taken to develop. It may take less than a hundred years, or many thousands of years. At any one location, it depends on the stability of the soil-forming factors or prevailing environmental conditions. When a soil reaches a dynamic equilibrium with its environment, it will change very little over time unless there is a change in one or more soil-forming factors. In southern Canada, where the last glaciation covered most of the land until about 10-12 000 years ago, fully developed soils are normally less than a meter thick. In contrast, beyond the glacial maximum in the USA or in areas glaciated in earlier times, soil thicknesses may be several metres. This has more to do with the changing conditions for soil formation than the time the surfaces have been exposed.

Within similar time intervals, soils in tropical areas may attain thicknesses of tens of metres (Brady and Weil 2002), whereas at the other end of the climate spectrum, Antarctic soils are generally only a few centimetres thick (Birkeland 1999), as are soils in extremely arid environments where there is little water to promote weathering (Fig. 1). Because the parent material progressively breaks down during the formation of soil, the soil mineralogy may also be helpful in determining the length of time it took to form the soil (Birkeland 1999). For example, the amount of clay, which is a normal byproduct of mineral weathering, is likely to increase over time, and can therefore be helpful in estimating the duration of the weathering process.

WHAT ARE PALEOSOLS?

In simple terms, a paleosol is:

- A weathered zone at the modern surface that formed under different environmental conditions; or
- A soil that has subsequently been buried by younger deposits and is no longer part of the active weathering process.

Paleosols are not always easy to recognize. For example, subsurface oxidized zones that are caused by groundwater percolating through fractures or joints in rocks, sands, or gravels are not paleosols because they do not develop the true properties of weathering zones such as horizons, which show a progressive downward change in mineral alteration, structure, etc. As indicated in Figure 2, the soil profile is an indirect reflection of climate, among other things, at the time of soil development; the more complete the profile, the more climate information can be derived.

Paleosols are found worldwide under a variety of conditions. However, a few places where extensive studies have been undertaken stand out. For example, paleosols formed in glacial deposits, such as the Sangamon paleosol in the United States and Canada (Hall 2000) and the Eemian soil in Europe (Stremme 1998), have been important in identifying ancient surfaces and interpreting interglacial climates. What makes glaciated regions unique is that distinct climates are commonly associated with specific material types, such as cold climates with till, and warm climates with superimposed soils (Fig. 3). It is worth mentioning that the early work on glacial stratigraphy in the mid-western USA, and to a lesser extent in Europe, was based largely on the recognition of buried soils within glacial sediments (Sibrava et al. 1986). However, areas covered by windblown deposits contain the most widespread and largest number of buried soils. The mid-western USA (Rutter et al. 2006), Alaska (Beget 2001), the north-central region of China (Rutter et al. 1991; Rutter and Ding 1993; Fig. 4), eastern Europe (Stemme 1998), and central Asia

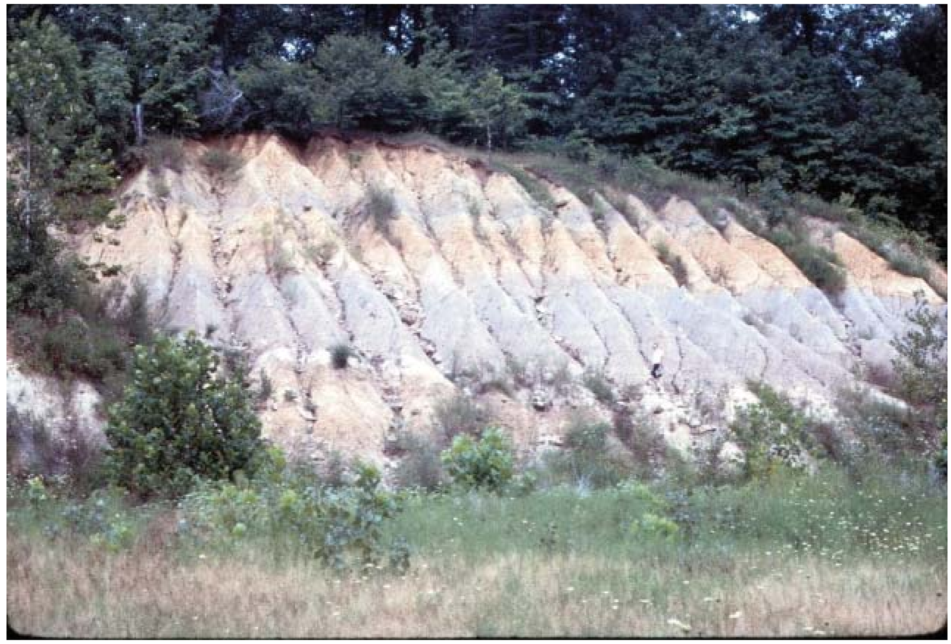

Figure 3. Photograph illustrating late glacial Wisconsinan silt (pink) overlying Illinoian glacial till (grey). The modern soil in the silt is superimposed on the Sangamon paleosol that has developed in the Illinoian till. The latter overlies a second till, the Kansan (grey), which hosts an older Yarmouth paleosol (pink). The Yarmouth paleosol in turn overlies a lower silt (pink). Although not clearly seen here, A, B, and C horizons are present in both paleosols (see Hall 2000). This section is about 8 m thick and is located near Cagles Mills, Indiana, USA. Photo by J.T. Teller.

(Bronger et al. 1998) are all locations of thick loess units that contain paleosols. Interbedded paleosols are also commonly found in Holocene sand dunes throughout the world (e.g. Wolfe et al. 2000; Fig. 5); these paleosols are usually poorly formed because they were commonly covered by sand before they had time to fully develop. Multiple buried paleosols are also common in both ancient and modern floodplain deposits (Autin and Asian 2001), though usually they are not as abundant as in windblown deposits. Relict and exhumed paleosols are generally not as useful as buried paleosols in reconstructing paleoclimates because they may have a history of superimposed or polygenetic soil development.

INTERPRETING CLIMATE CHANGE FROM PALEOSOLS

Although all soil-forming factors play a role in what type of soil may develop over time, moisture and temperature are particularly important. A buried soil, relict soil, or exhumed soil may have been subjected to hundreds, thousands, or millions of years of exposure and to a variety of pedologic changes, including a number of different climatic conditions. Therefore, using the Order or Great Group of the original soil to interpret the climate in which it formed (e.g. Fig. 2) can be challenging, because changing climate may result in a new soil being superimposed on the original profile. Usually, a diagnostic horizon of the dominant soil type can be identified, and one can conclude that conditions conducive to the development of that horizon existed for some interval of time, which may be enough information to interpret the original formative climate. In other cases, similar or dissimilar horizons in the same profile signify that more than one profile is superimposed on another, forming a polygenetic soil. The number of possible complications increases with soil age and the frequency of pedologic variations.

Buried paleosols are probably the most useful proxies for reconstructing past climates. However, as with relict and exhumed paleosols, there are problems that must be addressed. Ideally (but rarely), a pale-

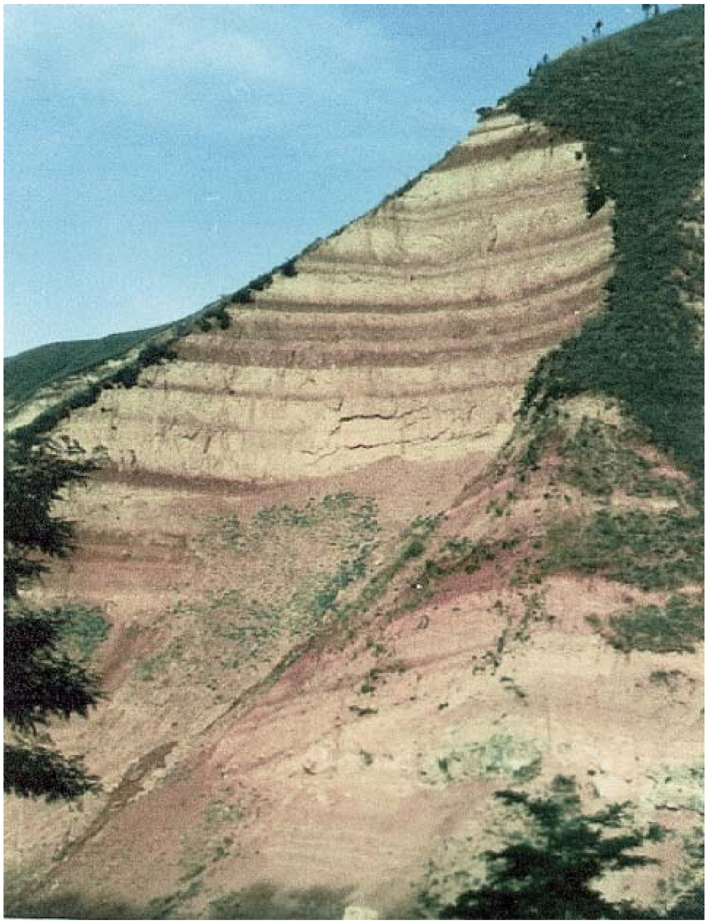

Figure 4. Photograph showing the loess–paleosol sequence located near Xian, near the southern part of the Chinese Loess Plateau, about 200 km east of Baoji. This section is similar to the section at Baoji described in this paper. The section shows paleosols (red) alternating with loess units (beige) that together represent over 2 Ma of earth history. The section is about 150 m thick.

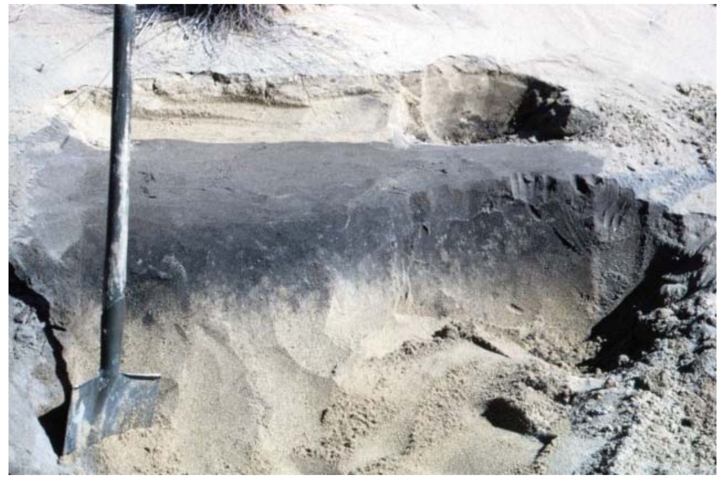

Figure 5. Photograph of a paleosol consisting of an organic-rich zone (an A horizon) in Holocene dune sands, overlain by more recent dune sand. This section is about a metre thick (note the shovel) and was taken in the Carberry Sand Hills, Manitoba. Photo by J.T. Teller.

osol exhibits all the characteristics of an equivalent modern soil. Commonly, however, the buried soil is polygenetic, or buried before it was fully developed, or contaminated with material that has been leached or filtered from above, thus altering the characteristics of the original soil and complicating identification, not to mention a paleoclimatic interpretation. Several characteristics or components of soils that may aid in the interpretation of paleosols are discussed in the following sections.

Mineral Horizons

As stated above, the more complete the buried profile, the easier it is to identify the original soil type prior to burial, and to predict aspects of the former climate. In most cases, the organic horizon is missing, and even the minerogenic horizons (A, B, C) may be partly missing because of erosion or incomplete development. However, if one or more of these horizons can be identified, a clue to the former climate may be elucidated. Commonly the details of individual horizons are particularly helpful in interpreting formative conditions, and these may be designated by a variety of suffixes or modifying letters added to the main horizon designation. The more 'suffixes' that can be identified, the more will be known about past climates. Below are a few of the more common horizons identified in Canadian soils that have been helpful in deciphering paleoclimate (see the Canadian System of Soil Classification (1978) for a complete listing of modifying suffixes).

i) An 'Ae' horizon is characterized by a downward removal (leaching) of soil material such as clay, Fe, Al, and organic material, and is usually found immediately above the B horizon;
ii) An 'Ah' horizon comprises that part of the A horizon that is normally rich in dark organic matter;
iii) A 'Bf' horizon is enriched in amorphous material, principally Al and Fe combined with organic matter;
iv) A 'Bm' horizon is a B horizon that is slightly altered by hydrolysis, oxidation, or solution; it is characterized by weathering and structural modification of parent materials, and partial or full removal of carbonates;
v) A 'Bt' horizon contains accumulated clay, as indicated by finer soil texture and by clay layers (cutans) coating soil particles or aggregates (peds) and lining pores; prismatic or columnar structure is usually developed;
vi) A 'Bw' horizon is designated if not enough clay has accumulated for the soil to be called a Bt. Bw is used in the USA system (National Soil Survey Center (NSSC) 1995) and when describing certain paleosols found in the Chinese Loess Plateau (see section on the Paleosols of the Chinese Loess Plateau, page 73); and
vii) 'Cca' and 'Ck' horizons are similar, both indicating certain percentages

of secondary calcium and/or magnesium carbonate above the amount contained in unenriched parent material.

These and other soil horizons are identified by a variety of criteria that must be met before a designation can be made for a certain soil Order, Great Group or Subgroup. Other than some A horizons, B horizons in paleosols are probably the most diagnostic and helpful mineral horizons in identifying original formative conditions. Some Orders may have the same basic type of B horizon, but with variations. For example, a Luvisol and Chernozem may each contain a Bm horizon, but the Bm in the Luvisol has different characteristics from those seen in the Bm of a Chernozem. The exact requirements for a specific horizon of a Canadian soil can be found in the Canadian System for Soil Classification (1978) or for other countries in the soil classification applicable to that country (e.g. in the USA, NSSC 1995). Other problems may be encountered, such as post-burial changes that mask or alter the original characteristics of the soil. For example, aqueous solutions from above may leach salts, which then precipitate in the paleosol, altering the pH and Eh and thus changing the physical characteristics of the original B horizon.

Organic Matter

Every actively forming soil contains organic matter (Brady and Weil 2002). Soil organic matter is the major source of carbon in a soil, exceeding CO_2 and soil carbonates. In most soils, organic matter decreases and approaches zero with depth, although the depth varies. For example, the organic content of grassland soils is higher, and is found to greater depths than in forest soils because of the greater root density and slower rate of decomposition in the former. Most organic compounds in a soil are passive (i.e. not biologically active), as opposed to the active fractions, which are characterized by microbial activity and decomposition. The passive fraction is stable and achieves a steady state within about 5000 years (Birkeland 1999). The long-term net accumulation of soil organic carbon varies with type of vegetation and is determined by the balance between accumulation and destruction: boreal and temperate forests have comparable net accumulation rates (10-12 grams of carbon/m^2/year), followed by tundra sedge moss (6 g C/m^2/year), polar desert and temperate grassland (2 g C/m^2/year), and finally temperate desert (< 1 g C/m^2/year; Schlesinger 1990).

The amount of organic matter remaining in a paleosol is influenced by climate, drainage, and vegetation type; normally, organic production (hence, organic content) in soils is higher in cool moist regions, in poorly drained soils, and where root biomass is greatest. Buried soils may have the organic part of their A horizons missing or only partially preserved. If they are preserved, they can be used with other criteria to identify a soil to perhaps the Subgroup level. For example, a Chernozem with a brown Ah horizon suggests a dry climate having a grassland cover, whereas Ah horizons found in Regosols in young sand dune deposits indicate rapid burial and a dry climate. If part of an A horizon is present in the profile, the lower horizons may be intact, and a fairly precise identification of the original soil is possible. A useful biogenic component found in A horizons is plant phytoliths, which are relatively resistant, microscopic particles of opal (silica) or calcium oxalate that are found in many plants, and take many forms depending on the plant structure (Meunier and Colin 2001). Phytoliths are most useful in identifying vegetation type, specifically in elucidating the relative amount of grass cover as opposed to forest cover, allowing speculation on past climate.

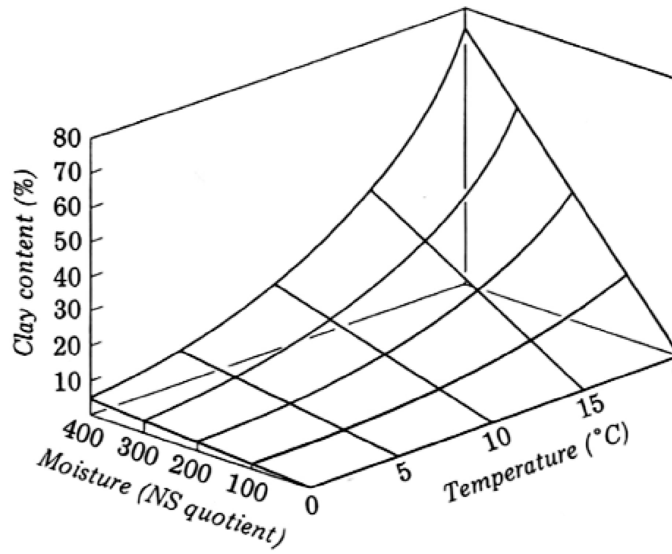

Figure 6. Idealized relationship between clay content (in the top ca. 1 m of a soil derived from igneous rocks), moisture, and temperature. Modified from Jenny (1935).

Clay Content and Clay Mineralogy

The clay content of a soil may originate from parent material, from surface influx, and/or by pedogenic processes, such as weathering and translocation within the soil profile. In general, the amount of clay produced by pedogenic processes is greater where conditions are hot and wet than in regions that are cold and dry (Jenny 1935; Fig. 6). For example, if a buried soil has characteristics similar to a Chernozem, but has an unusually high content of clay in the B horizon, it could mean that it was formed under wetter conditions, as more clay was translocated downward by weathering, or was produced *in situ* under more intense weathering.

Clay mineralogy is as important as clay content in aiding the interpretation of paleosols. Although clay minerals may be present in the original igneous, metamorphic, and sedimentary rocks, the original clay minerals in these materials may be altered by later weathering. The type of clay mineral that is formed during weathering is largely controlled by water chemistry, amount of rainfall, and rate of leaching, and, therefore, relates indirectly to climate. Clay minerals generated from parent material that is low in primary clay content probably constitute clay mineral assemblages that are stable in that weathering environment. On the other hand, soils formed from parent materials with a high content of inherited clay minerals may have unstable clay mineral assemblages that gradually change to more stable forms. The change will progress from the surface downward, and complete change to stable assemblages may take considerable time, depending on soil water chemistry (Birkeland 1999). It is apparent that when using clay mineralogy to identify climate change from a paleosol, one needs to compare the paleosol with the modern soil, other paleosols, or the parent material. The greater the contrast, the easier it is to draw conclusions about climate changes. For example, if the modern soil is a Podzol from a cool and wet mountain setting that is poor in highly-weathered clay minerals such as kaolinite, but the buried soil in the region has a high percentage of kaolinite, then the buried soil may have formed under at least a warmer climate and perhaps a wetter climate than the Podzol.

In summary, clay content and clay mineralogy in paleosols are useful in interpreting climate and climate change. However, clay mineralogy is complex, and clay minerals can originate in several ways, so analyzing a single paleosol in order to gain climate information can be difficult. Comparing the paleosol with soils known to have formed under specific conditions, such as the modern soil, is important.

Red Soils

Redness of soils, in conjunction with other data, can yield information on climate (Birkeland 1999). For example, in parts of California, New Mexico, New Zealand, and Colorado, older soils are redder than younger ones, apparently because they required, among other climatic factors such as warmer and/or wetter conditions, thousands of years to reach equilibrium with the current climatic conditions before they acquired the red colour. Furthermore, factors other than time and climate (temperature and precipitation) may influence soil redness, such as the type of Fe-bearing minerals in the parent material, the rate of leaching, and redox potential. Red soils form relatively quickly in areas of high temperatures and little precipitation, such as deserts, but form somewhat more slowly in wetter, humid, tropical regions (Blodgett et al. 1993). The rate of increase in redness is very slow in cold, Arctic and Antarctic regions. In addition, redness in paleosols may mask later soils development and hinder their interpretation.

As can be seen from the foregoing, redness in soils may have a complex history, so interpretation must be used with caution. If red B horizons are found in buried soils and the modern soil is different, the redness should aid in interpreting past climate. In general, weathering in wet tropical areas tends to produce red soils such as Laterites (Oxisols). However, there still may be a question as to whether these red soils formed under rigorous climatic conditions, or are the product of extended pedogenesis under a relatively stable climate. In Canada, red soils and red paleosols are not common.

Soluble Salt Accumulations

The most common soluble salts in soils are calcium and magnesium carbonates and calcium sulfate (gypsum), although other salts such as halite occur in some soils. Precipitation is most important in determining the accumulation of soluble salts because infiltrating water in wet climates normally leaches the most soluble components from the weathering zone. Soil permeability also plays a role in this. The type and amount of soluble salts in a soil depend mainly on the presence of these ions in the parent material and additions from the atmosphere. In some cases, such as where calcite accumulations occur in a soil (e.g. Cca horizons), groundwater may also contribute the needed ions. In most cases, in order for soluble salts to form in the lower soil zone, ions must be leached downward from the A and B horizons. The C mineral horizon typically has letter suffixes representing various secondary minerals added such as calcite, hence the Cca or Ck designations in Canadian soils (Fig. 2). The depth to the top of the C horizon depends on the intensity and length of time of the weathering processes. Calcretes (caliche, hardpan, duricrust) are formed mainly in semiarid and arid regions and consist of hard, dense deposits of secondary carbonates that occur as nodules, lumps, crusts, or thick concrete-like layers in the lower soil zone. Calcretes may be formed by capillary rise and evaporation of $CaCO_3$-charged groundwater. In places, where the overlying A and B horizons have been eroded, calcretes lie at the modern surface, and may form resistant caps on mesas in arid and semiarid regions. Field identification and description of a buried paleosol to the Subgroup may be hampered if leaching and evaporation have deposited salts that overprint the original profile.

Polygenetic or Superimposed Paleosols

The presence of paleosols (exhumed, relict, or buried) provides an opportunity to interpret former surfaces, despite many challenges. For example, how many of the soil-forming factors, especially climate, changed during this interval of exposure, and were there different environmental conditions (e.g. climate) through time that led to the development of superimposed soil characteristics on the original profile? In general, the shorter the time a surface has been exposed, the less likely it is that multiple generations of soils related to different controlling conditions such as climate will be superimposed on the landscape, and the easier that soil is to interpret. Usually, in a polygenetic profile, the most highly weathered or best-developed paleosol can still be identified, although the best developed soil may not reflect the original profile; i.e. earlier or later soil development will not mask the better-developed soil. For example, where a Laterite has formed, its characteristics will not easily be masked or obliterated by subsequent conditions that favor formation of a Podzol, which forms under less-rigorous climatic conditions. Where a change in formative conditions does impact on a previously developed soil, multiple horizons may aid in identifying multiple paleosol-forming conditions. For example, two distinct Bm horizons separated by a C horizon, or a slightly weathered parent material, suggests two periods of Chernozem development under different environmental conditions.

ESTIMATING THE AGE OF PALEOSOLS

The length of time that a surface has been exposed to weathering can often be determined, although there may be uncertainties. In a general way, in one climatic region and with similar parent material, thick soils are older than thin ones. Once a soil is buried, and away from the influence of weathering at the interface with the atmosphere, soil formation ends. Estimating the length of time it took for any soil to develop, i.e. before it ceased to actively form because of burial, is difficult to calculate, in part because a soil typically forms over many thousands of years. When the soil was buried and for how long it has remained buried depends on the age of the overlying material – an entirely different story than the time it took for the paleosol itself to form. Late Quaternary paleosols contain both organic and inorganic carbon and may lend themselves to radiocarbon dating, although the method is limited to about 35 to 40 000 years and is fraught with complications. For example, as a soil develops, both old and new carbon is added, exchanged and mixed throughout the profile, making the carbon system very complex. The average time that carbon spends within a given part of the soil profile is called the mean residence time (MRT). Various fractions of soil organic matter yield different MRTs, which is a reflection of the stability of the individual fractions (Birkeland 1999). The MRTs commonly increase with depth in various types of soils, but of course, the MRT has little meaning in surface soils, because organics are continuously being added. A model to estimate MRTs in soil sequences is presented in Figure 7 (Wang et al. 1996). The radiocarbon dates are predictably much less than the estimated age of the soils themselves, because organic carbon continued to be added to the profile until burial, merging with older carbon that was added from previous root systems. Organic matter at the top of a buried A horizon provides the maximum age of deposition of overlying deposits, whereas dating the deepest pedogenic organic matter yields the minimum age of the deposits in which the soil formed. Mixing or addition of material such as wind-transported carbon can also alter radiocarbon ages (Birkeland 1999). In addition, the mobility of various components of organic mater, and the organic fraction used for the age, such as humus and/or humic acids, can result in different ages at different levels of the horizon being dated.

Radiocarbon dating of inorganic carbon, mostly secondary calcium carbonate, is another method of establishing the age of a paleosol. Problems occur because $CaCO_3$ is extremely soluble and can be deposited, put into solution, and re-deposited in a variety of circumstances during the period of soil formation. Additionally, carbon of any age can be added from the CO_2 in the soil; hence, organic carbon can contaminate the inorganic fraction. Datable secondary inorganic carbonate can be deposited on soil particles, or even on the base of pebbles below the main soil horizons. The inner, older layers of carbonate in these deposits are protected by younger layers, offering the best opportunity for determining maximum soil age. Secondary carbonate in the lower part of a C horizon, where solutions may no longer be active, also offer opportunities for finding the oldest age. If there is no exchange of carbon with an external source after pedogenic carbonate formation, then a date may be relatively accurate (Amundson et al. 1994), although these conditions are uncommon, and accurate dates using inorganic carbon are therefore difficult to achieve.

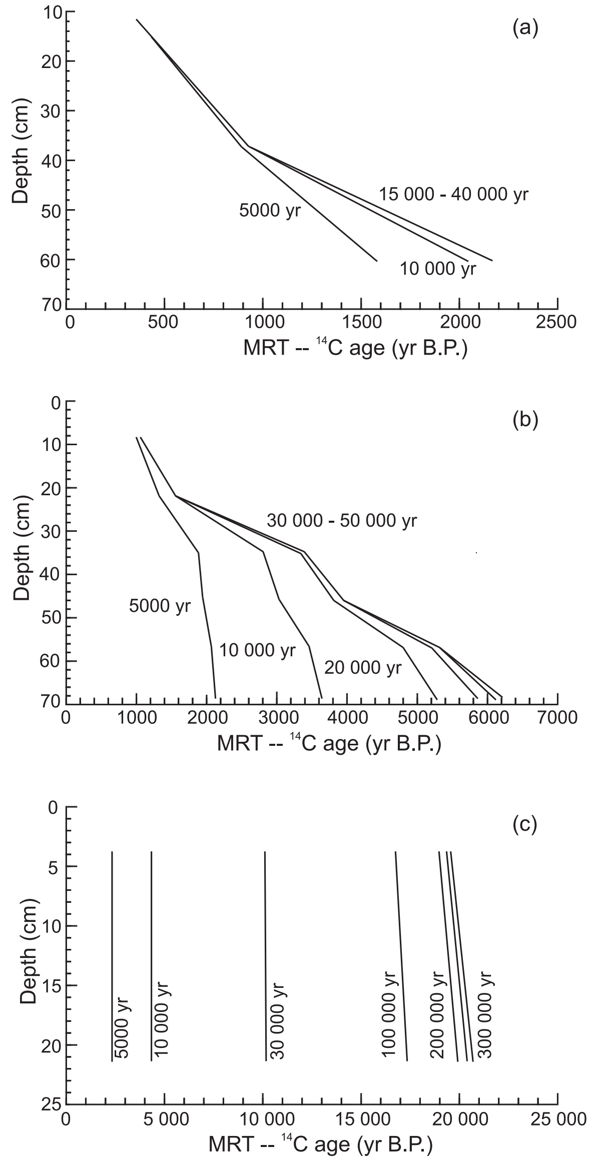

Figure 7. Evolution of ^{14}C age (MRT) of soil organic matter for a) a forest soil; b) a prairie soil; and c) a desert soil. Different curves represent the calculated ^{14}C ages of soil organic matter at different times of the soil's development. Numbers along the curves indicate the true age of the soil (Wang et al. 1996).

Another method that has recently been applied in dating paleosols, with varying success, is the uranium-series disequilibrium method, which has been tried on the innermost coating of secondary carbonate precipitated on pebbles in the lower soil zone. Results are mixed because of contamination from externally sourced uranium (e.g. in groundwater) and because of the uncertainty concerning when clasts originally became coated, which can range from a few thousand years to hundreds of thousands of years (Slate et al. 1991). Sharp et al. (2003) claim reliable and precise U-series dates on pedogenic carbonate rinds on pebbles from gravels in glaciofluvial terraces in Wyoming. These ages were obtained by $^{230}Th/U$ thermal ionization mass spectrometry (TIMS) dating applied to carefully selected milligram-size samples.

The most accurate way of determining the age of a paleosol, that is, the amount of time needed to develop the soil itself, is to date the material that overlies and underlies the paleosol. For example, if the material overlying the paleosol near the soil contact is 10 000 years old and the underlying material near the lower contact is 15 000 years old, then the soil took about 5000 years to develop.

Depending on what is dated, a variety of other techniques may be applicable; however, indirect dating of paleosols by analysis of ^{26}Al and ^{10}Be cosmogenic nuclides in quartz from buried soils (Balco et al. 2005a, b) has become more common in recent years. This method involves measuring nuclide concentrations at multiple depths in the paleosols. The geologic context of the sample is used to construct an exposure/burial history that can be used to predict the nuclide concentrations in the samples as a function of unknown parameters such as ages of geological units. An exposure model is developed that results in the best fit between predicted and observed nuclide concentrations. This method has successfully dated till and loess units from paleosols in the older Quaternary deposits in the central USA (e.g. Balco and Rovey 2008).

THE PALEOSOLS OF THE CHINESE LOESS PLATEAU

Paleosol investigations in Canada have been undertaken by a number of researchers (see Dormaar 1978; Mahaney et al. 1987; Tarnocai 1990). Paleosols studied in northwest Canada include multiple buried soils between tills and relict soils developed in glacial deposits of varying ages (Rutter et al. 1978; Tarnocai et al. 1985; Hughes et al. 1993). In the Canadian Cordillera, investigations have been carried out on buried soils within early Pleistocene multiple tills in the southern Rockies (Karlstrom and Barendregt 2001) and late Pleistocene and Holocene tills in central BC (Alley et al. 1986; Karlstrom and Osborn 1992). In the Prairie provinces, buried soils have been studied in a wide range of environments, including Holocene loess and sand dunes (Fig. 5), Pleistocene till surfaces, and marine deposits (David 1966, 1993; Dormaar and Lutwick 1971; Turchenek et al. 1974; Mills and Veldhuis 1978; Waters and Rutter 1984; Vreeken 1993). In Central and Atlantic Canada, buried soils in several late Pleistocene inter-till and Holocene dunes have been studied (Karrow et al. 1982; Mahaney 1985; Rutherford 1989; Mahaney and Hancock 1990; McCann and Byrne 1994). However, the loess/paleosol sequences in the Chinese Plateau are often cited as the best case study, because the Loess Plateau represents the longest known continuous Quaternary paleosol record, and offers a large amount of information on climate change and the forcing mechanisms involved (Fig. 4).

The Chinese Loess Plateau is extensive, covering an area of 640 000 km^2 in north-central China (Rutter et al. 1996). Loess thicknesses vary between about 600 m in the northwest to about 200 m in the southeast. Deposits consist of silt with varying amounts of fine sand and clay. Today's climate in the northern part of the Loess Plateau is arid (<200 mm annual precipitation) becoming more temperate toward the south (>800 mm annual precipitation). These conditions persisted with some modifications throughout the entire Quaternary Period and possibly longer (Ding et al. 2001). Loess units alternating with superimposed paleosols reflect dynamic, long-term (millenia-scale) climate variations. The relatively thick, coarse grained units of loess were deposited during extended periods of cold, as dry winter monsoon winds carried dust from the north, northwest, and west (Rutter et al. 1996). The source of these silts and the dominant wind direction were essentially constant during the Quaternary, so the loess sequence has an overall grainsize decrease toward the south and southeast, with units thinning in the same direction. This distribution is (and was) a function of the winter monsoon and is controlled mainly by the position and intensity of the Siberian High Pressure System. Grainsize variations in the unweathered loess reflect varying wind strengths.

The interbedded soil horizons in the loess indicate periods when warm, moist, summer monsoons penetrated from the south. During these warm periods, the rate of loess deposition was reduced, or ceased altogether. This resulted in alternating units of loess and soils that developed over the last 2.6 Ma (Rutter et al. 1991, 1996). Determining the age of the various units was accomplished by radiocarbon, thermoluminescence and optical stimulation luminescence methods, sedimentation rates, paleomagnetic signatures, and an orbital time scale developed from the loess record (Ding et al. 1994, 2002).

The alternation of the loess–paleosol sequences is best explained by the Milankovitch theory of orbital forcing (Ding et al. 1994, 2001). From ca. 900 ka BP to the present, loess units are thicker and soils well developed, and the 100 000-year orbital eccentricity cycles dominated. From ca. 1.6 Ma to 900 ka BP, when the 40 000 year obliquity (tilt) cycle dominated, the loess units are relatively thin, and soils are less well developed. However, all orbital cycles, including the 19 000 to 23 000-year precessional cycle, were operative regardless of which climatic conditions dominated and what soil developed (Rutter et al. 1996). From 1.6 Ma to 2.6 Ma, the picture is less clear, and it appears that no single forcing mechanism dominated.

The climatic variations recorded in the Chinese loess sequence were interpreted using a number of climatic proxies, the most important being the type of soil, the texture/grain size of the loess units, and magnetic susceptibility. The Baoji loess–paleosol sequence described here is near the city of Baoji, about 200 km west of Xian in the southern part of the Loess Plateau. The section, about 165 m thick, contains at least 37 paleosols developed in 37 distinct loess units (Rutter et al. 1991; Rutter and Ding 1993; Fig. 8). Fortunately, polygenetic paleosols are rare, making identification and interpretation easier. Visually, the sequence consists of bands of red paleosols separated by beige loess units; thickness of individual units varies, but the appearance of each paleosol–loess unit is similar. However, close study has revealed important differences. Complete A horizons are common, but many have been partially or totally eroded, and in some cases part, or all, of the B horizon is missing. Further, carbonate accumulations and nodules may or may not be present. Field investigation has helped to identify the master horizons, but labo-

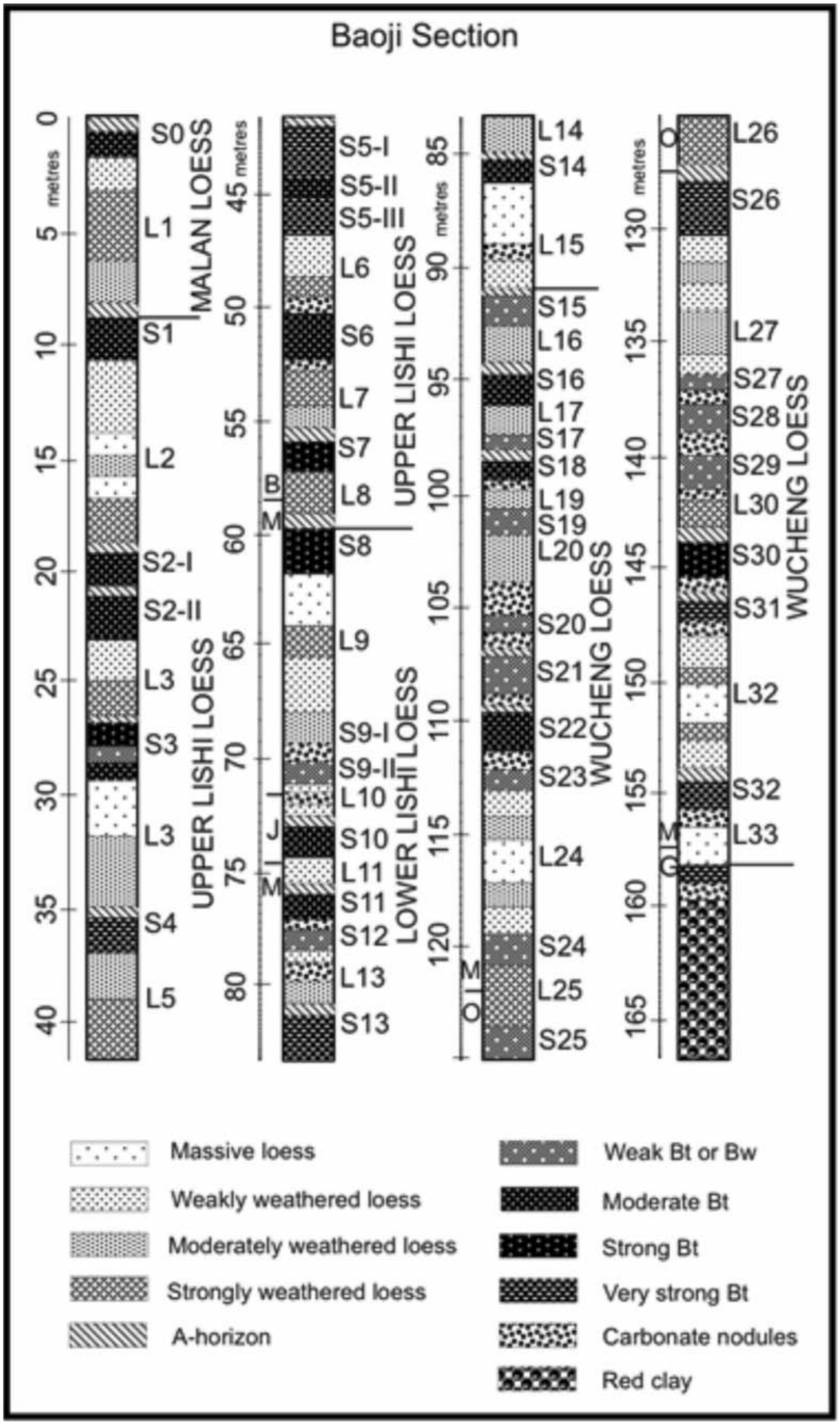

Figure 8. Loess–paleosol sequence at Baoji with major paleosol horizons indicated by the letter 'S'. Paleomagnetic reversals: B = Brunhes; M = Matuyama; J = Jaramillo; O = Olduvai; G = Gauss. L1 to L33 – loess intervals; S1 to S32 – paleosol intervals. Modified after Rutter and Ding (1993).

ratory analyses, especially the micromorphological characteristics of the paleosols, were predominantly used to determine pedological processes and diagenesis, and ultimately to interpret paleoclimate (Rutter and Ding 1993). Although organic accumulation and decomposition was a vital process during the development of the soils and of most of the loess, organic material remains in only few loess and soil units, and therefore cannot be used effectively in climatic interpretation.

Primary carbonate in all paleosols has been extensively leached. Secondary carbonate, however, is precipitated mostly below the B horizons, forming Cca horizons that are most abundant below the L9 loess unit (Fig. 8). Clay translocation is the most common pedological process in the development of many of the soils, and can readily be seen in the field and in thin section. This process forms the Bt horizon, which is by far the most important horizon in identifying and interpreting past climates from the paleosols (Rutter and Ding 1993).

Only in the upper part of the sequence, above loess unit L9, do paleosols commonly consist of a relatively strong A-Bt-C sequence, whereas below L9 a relatively weaker sequence is more common (in this context, 'strong' and 'weak' denote the relative degree of development of a given soil horizon or horizons). In the field, the more well-developed soils display clay coatings on ped surfaces of the Bt horizons, particularly in the soils above L9. In some soils in the middle part of the section, however, clay coatings appear less abundant, indicating less well-developed soils.

By noting the presence or absence of gleization (a process involving the reduction of iron in saturated soils, commonly indicated by black Fe-Mn films and mottlings), and the concentration and character of clay coatings, it is possible to recognize four types of B horizons. These include a weak Bt or Bw, a moderate Bt, a strong Bt, and a very strong Bt; the specific type that is present is a reflection of the intensity of weathering, allowing the climate and vegetation under which they formed to be interpreted. The strongest Bt horizons suggest the presence of a relatively dense forest or steppe–forest vegetation, and a relatively stable landscape with a pronounced seasonal wet–dry alternation during soil development; these soils are dominant in the upper part of the sequence from near the Brunhes/Matuyama boundary (B/M in Fig. 8), although there are also several units with strong Bt horizons in the lower part of the sequence (Figs. 8, 9). In contrast, at the other end of the spectrum, a weak Bt horizon in the buried soil represents a steppe–forest vegetation, a relatively stable landscape and wet–dry alternation, but with generally less precipitation than that required for a strong Bt; these are most common below the B/M boundary (Fig. 8). In Canada, this soil spectrum is indicated by luvisols at the wet end and chernozems at the dry end. The development of Chinese paleosols occurred under a climatic regime similar to that prevailing in the Loess Plateau today, where weathering occurs within the context of the humid summer monsoon and the dry-cold winter monsoon. Figure 9 shows the climatic oscillations over the past 2.6 Ma and the degree of soil weathering.

CONCLUSIONS

Paleosols mark old surfaces that may have been in existence for hundreds, thousands, or millions of years, and have been a valuable tool in distinguishing Quaternary events and interpreting past environmental conditions. Buried soils are also increasingly being used in interpreting the older geological record (Ratallack 2001). Buried soils may be easier to interpret than relict or exhumed soils, because the pedologic/weathering history may be simpler; i.e. the buried soil may not have undergone as many climatic variations before it was buried as did a relict or exhumed soil.

Recognizing a specific B horizon is particularly valuable in aiding the interpretation of former climates. B horizons are usually diagnostic of the entire soil, although other horizons, especially the A, may be diagnostic for some soil Groups. Hindering recognition of diagnostic climatic indicators in paleosols, especially in relict paleosols, are superimposed or polygenetic profiles developed when climates or other environmental factors were altered.

Buried soils and some exhumed soils may have been affected by downward leaching of overlying materials into the paleosol, and may even have been accompanied by translocation of 'foreign' materials after the main soil-forming process ended. This would have altered the original characteristics of the paleosol, thus increasing the difficulty of identification and interpretation. Nevertheless, information on past climates can generally be derived from paleosols, and may include proxies for both climate and vegetative cover during weathering. Estimating the length of time it took to form a soil is based on the relative thickness of the solum in any given area; i.e. thicker soils take longer to develop than thinner ones. Determining the actual interval of time during which a paleosol formed is difficult. Although various methods can be employed, probably the most accurate, in the case of buried soils, is obtaining bracketing ages from datable material above and below the paleosol.

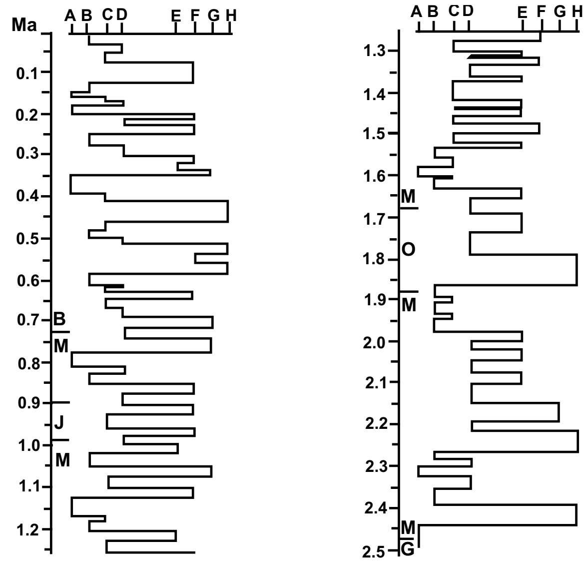

Figure 9. Climatic oscillations in the last 2.6 Ma interpreted from the Baoji loess–paleosol sequence, with wetter and warmer conditions increasing toward the right. A is associated with massive loess, B with weakly weathered loess, C with moderately weathered loess, D with strongly weathered loess, E with a weak Bt, F with a moderate Bt, G with a strong Bt, and H with a very strong Bt. Paleomagnetic reversals: B = Brunhes; M = Matuyama; J = Jaramillo; O = Olduvai; G = Gauss. Modified after Rutter and Ding (1993).

Quaternary paleosols are found in many parts of the world, and buried soils are particularly widespread in loess and alluvial deposits. The extraordinary loess-paleosol sequence in the Chinese Loess Plateau has not only revealed over 37 major climate cycles over at least the past 2.6 Ma, but also paleo-wind directions and insight into the climate-forcing mechanisms involved. In Canada, buried soils have been found mainly in glacial, alluvial, and eolian sequences. Especially useful are those found in multiple till sections in the Northwest Territories, and in glacial deposits of various ages in Yukon.

ACKNOWLEDGEMENTS

I thank Bill Mahaney and Dean Rokosh for their critical reviews, and Jim Teller and Geoscience Canada editorial staff for reviewing and improving the manuscript. Special thanks to Barbara Tilley for her help in preparing several figures.

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Submitted June 2008
Accepted as revised February 2009