An Evaluation of Methods for Identifying and Interpreting Buried Soils in Late Quaternary Loess in Alaska

Daniel R. Muhs

Thomas A. Ager

Josh M. Been

Joseph G. Rosenbaum

Richard L. Reynolds

Geologic Studies in Alaska by the
U.S. Geological Survey, 1998
U.S. Geological Survey Professional Paper 1615

An Evaluation of Methods for Identifying and Interpreting Buried Soils in Late Quaternary Loess in Alaska

By Daniel R. Muhs, Thomas A. Ager, Josh M. Been, Joseph G. Rosenbaum, *and* Richard L. Reynolds

Abstract

The presence of buried soils in Alaskan loess is controversial, and therefore criteria for identifying buried soils in these deposits need to be evaluated. In this paper, morphologic and chemical criteria for identifying buried soils are evaluated by studying modern soils developed mostly in Holocene loess under tundra, boreal forest, and transitional coastal-boreal forest vegetation in different parts of Alaska. Data from modern Alaskan soils that developed under vegetation similar to that of the present indicate that soil morphology, organic-matter concentrations, and P concentrations can be useful diagnostic tools for identifying buried soils. Soil morphologic criteria, particularly horizon colors and horizon sequences, are essential for identifying buried soils, but some minimally developed soils may resemble organic-rich alluvial, colluvial, or lacustrine deposits. Organic matter and total P contents and distributions can aid in such studies because in well-drained soils these constituents show rapid declines with depth. However, neither of these techniques may work if the upper genetic horizons of buried soils are eroded.

If buried soils are present in Alaskan loess, it would also be desirable to have techniques for determining the dominant vegetation under which the soils formed. Such techniques could then be used to reconstruct former vegetation types and paleoclimates in Alaska. A previous study suggested that tundra and boreal forest vegetation have distinctive carbon isotopic compositions, although both are dominated by C_3 plants. If this is the case, then the carbon isotopic composition of organic matter in buried soils could be used to reconstruct former vegetation types. A larger suite of modern soils from Alaskan tundra and forest were analyzed to test this hypothesis. Results indicate that modern soil O horizons in these two biomes have the same range of $\delta^{13}C$ values, and therefore carbon isotope compositions cannot be used to reconstruct former tundra or boreal forest.

Introduction

Interest in loess stratigraphy has increased greatly over the past decade because loess sequences contain detailed records of Quaternary environmental conditions and could be the closest terrestrial equivalent to the oxygen isotope record in deep-sea sediments (Hovan and others, 1989). In Europe and the North American Midcontinent, most loess deposits accumulated during glacials or stadials, whereas buried soils developed during interglacials or interstadials. In China, loess records are more complex because loess is deposited during both glacials and interglacials. The same is true for Alaska, as loess deposition took place during the last glacial period but continues today (Péwé, 1975).

There have been conflicting interpretations of loess stratigraphy in Alaska. Although Alaskan loess has been studied for decades (see summary in Péwé, 1975), prior to the 1980's there was little or no mention of buried soils in these deposits. More recent studies, however, report the existence of numerous buried soils in Alaskan loess sections, suggesting that episodes of loess fall alternated with periods of landscape stability characterized by little or no loess deposition (Hamilton and others, 1988; Begét and Hawkins, 1989; Begét and others, 1990; Begét, 1990, 1996; Hamilton and Brigham-Grette, 1991; Muhs, Ager, and others, 1997; Muhs and others, in press). Nevertheless, Péwé and others (1997, p. 29–34) question the existence of buried soils in the loess record of interior Alaska, and McDowell (1997) points out some of the difficulties in interpreting organic-rich zones in Alaskan loess sections. In order to resolve these conflicting interpretations, it is essential to establish which criteria can be used to identify buried soils in the Alaskan loess record.

Recognition of buried soils is often difficult. Birkeland (1999) outlines some of the field and laboratory criteria that are potentially useful, but many of these rely on the presence of well-developed subsoil Bt or Bk horizons, which are extremely rare in present-day Alaskan soils (Rieger and others, 1979). Organic-rich, surface-soil O and A horizons are found in Alaskan soils, and the challenge is to find methods of using these horizons for the identification of buried soils in the Alaskan loess record. However, organic-rich alluvial, colluvial, or lacustrine deposits can be mistaken for buried soil O horizons and vice versa. In addition, if the O or A horizons are removed by erosion prior to burial, it may be extremely difficult to identify the remaining parts of a buried soil.

In order to address these problems, it is necessary to characterize modern loess-derived soils that developed in diverse climatic and vegetation settings in Alaska. One purpose of this paper, therefore, is to evaluate methods for identifying buried soils in the Alaskan loess record. If it can be determined that key soil properties show depth-related changes caused by pedogenesis, then identification of buried soils is possible. Morphologic and chemical data are evaluated for modern Alaskan loess or loess-and-till-derived soils that are potentially useful in identifying buried soils in loess sections.

A second purpose of the paper is to evaluate a method for interpreting past vegetation types by looking at buried soils. Certain soil properties reflect the vegetation type under which they form, and therefore such properties in buried soils can be keys to past vegetation types (Jenny, 1980; Birkeland, 1999). Because much of Quaternary climate change in Alaska results in shifts in the dominance of boreal forest versus tundra (Ager, 1983; Ager and Brubaker, 1985), buried soils in loess sections could record these different vegetation regimes. However, as will be shown later, morphologic criteria alone may not be sufficient to identify the type of vegetation under which a buried soil formed. Carbon isotopes have been employed extensively for reconstruction of past vegetation types from buried soils (Cerling and others, 1989; Holliday, 1995; Fredlund and Tieszen, 1997; Muhs, Stafford, and others, 1997). This method was tested as a tool for distinguishing soils developed under boreal forest and tundra.

Methods

Modern Alaskan soils, developed in either loess or loess over till of either Holocene or latest Pleistocene age, were investigated in three climatically and biogeographically distinct parts of Alaska (figs. 1, 2): (1) Gelisols (Histoturbels, Haploturbels, Aquorthels, and Historthels) and Inceptisols

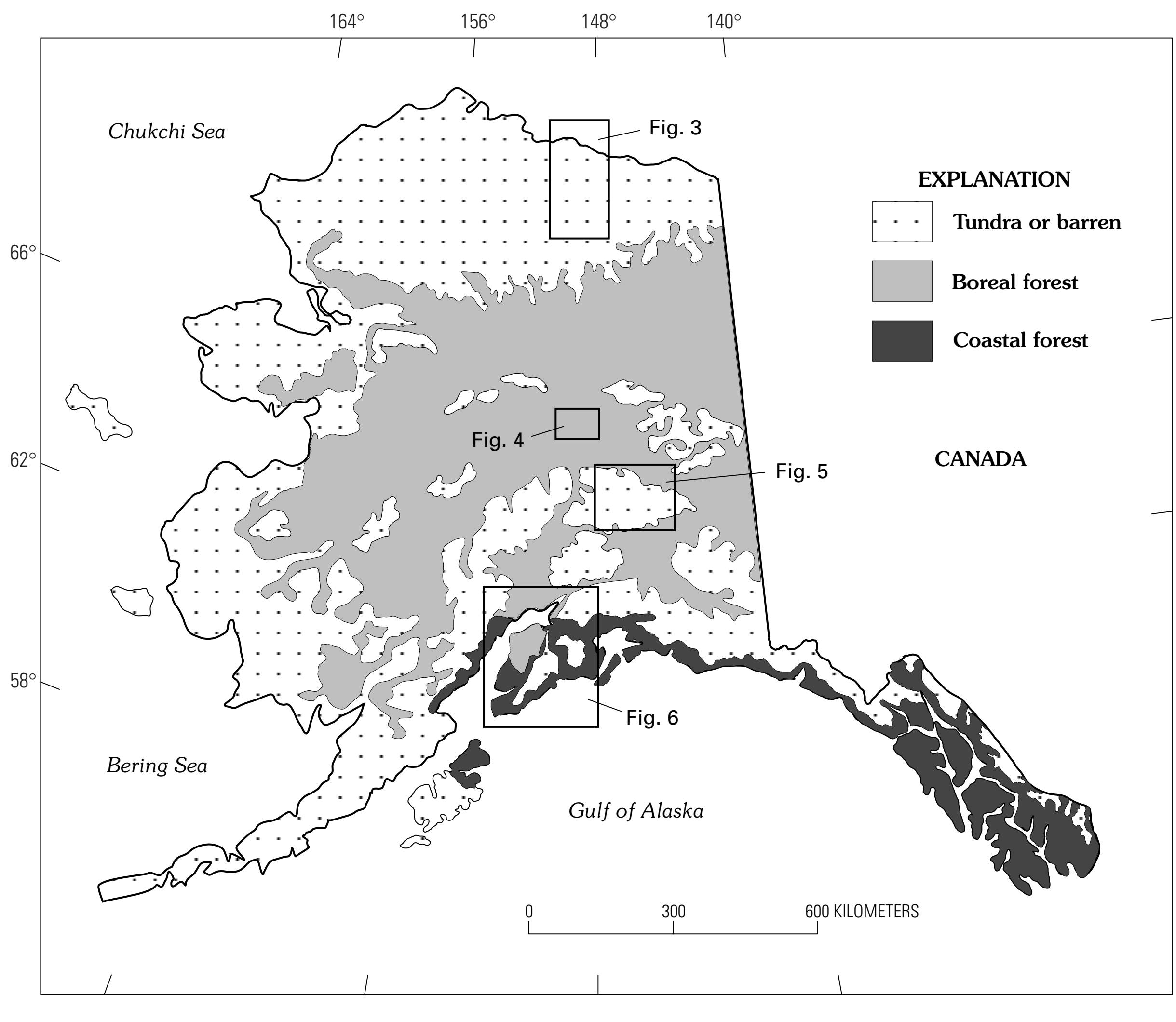

Figure 1. Map showing study areas and the present distribution of vegetation communities in Alaska (from Viereck and Little, 1972, as modified and redrawn by Ager and Brubaker, 1985).

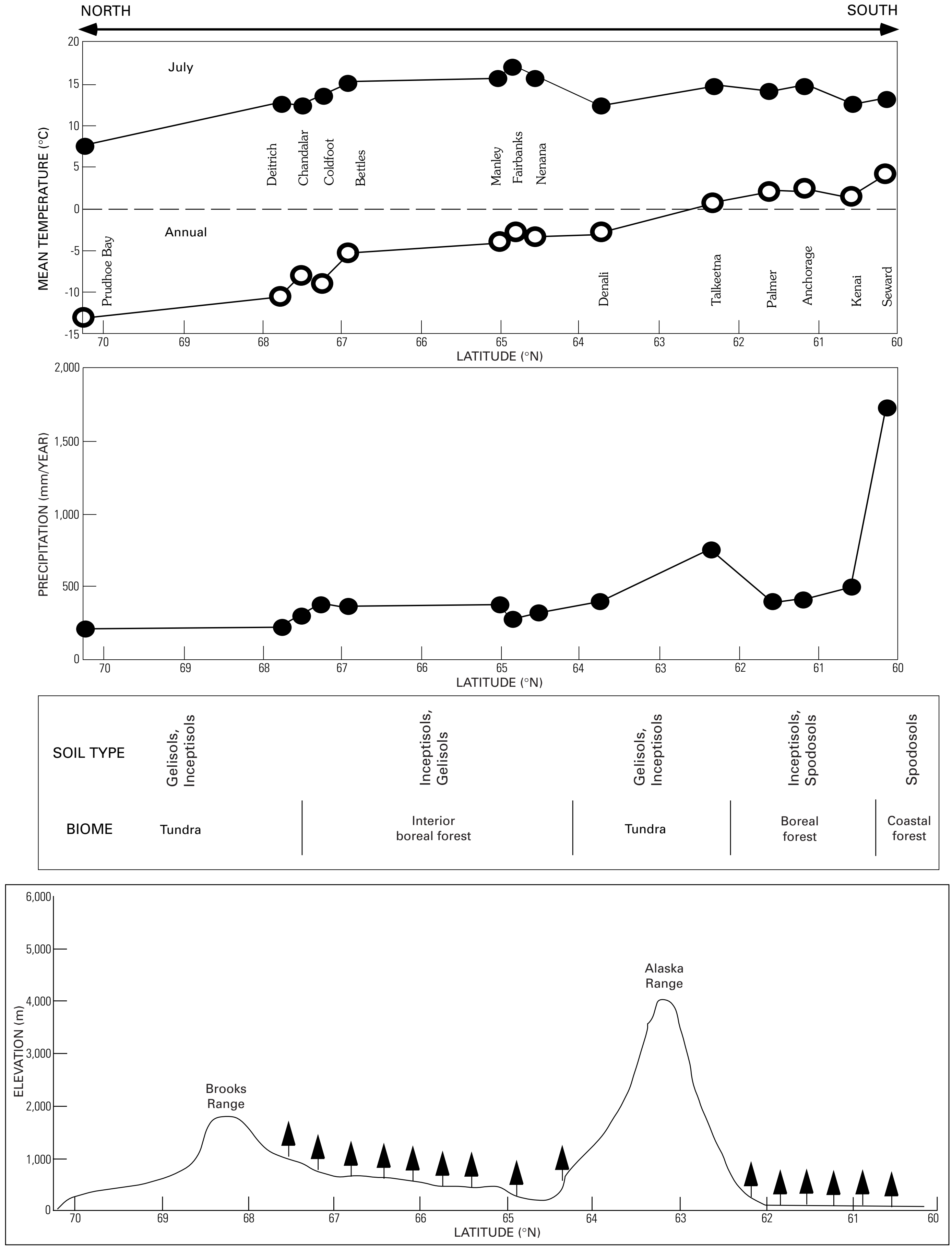

Figure 2. Transect of climates, major soil types, biomes, and topography from north to south across Alaska. Climate data are 1961–1990 means, except for localities where only shorter records exist (Prudhoe Bay, Deitrich Camp, and Coldfoot).

(Typic and Aquic Dystrocryepts) developed under moist or wet-moist tundra vegetation of the extremely cold, dry environment of the Arctic Coastal Plain, the Arctic Foothills, and the Alaska Range; (2) Inceptisols (Typic and Aquic Dystrocryepts) developed under boreal forest in interior Alaska near Fairbanks, where mean annual temperatures are higher and precipitation is greater than in tundra regions; and (3) Spodosols (Typic Haplocryods) developed under either boreal forest or transitional coastal-boreal forest near Kenai in southern Alaska, where precipitation and mean annual temperatures are higher than in the interior (fig. 2).

Organic-matter concentrations were estimated by weight loss on ignition at 950°C for 1 hour, slightly modified from the methods described in Ball (1964) and Nelson and Sommers (1982). As pointed out by Davies (1974), Dean (1974), and Nelson and Sommers (1982), some of the weight loss by this method is derived from structural water loss in clay minerals and CO_2 loss from carbonates. However, Nelson and Sommers (1982) note that estimation of organic matter content by the classical Walkley-Black method requires corrections both for unoxidized organic carbon and for conversion of organic carbon values to organic matter values. Janitzky (1986) reported a high correlation between organic carbon measured by the dry-combustion and Walkley-Black methods. We compared loss-on-ignition and Walkley-Black methods for soils near Fairbanks that have organic matter contents of less than 7 percent (as estimated by the Walkley-Black method). Results show that the relation between the two methods is linear with a high degree of explanation (r^2=0.73). However, loss-on-ignition values are consistently higher than Walkley-Black-derived values, even though the samples are not calcareous, suggesting some structural water loss from clay minerals. Therefore, loss-on-ignition estimates, of organic matter content less than ~10 percent reported here, should be interpreted as maxima. Certain soils studied on the Arctic Coastal Plain are calcareous; organic-matter estimates of these soils, derived from loss-on-ignition, are certainly maxima. Based on the estimated carbonate content, however, the error introduced does not affect interpretations.

Concentrations of P_2O_5, which are examined in some detail here, were determined by wavelength-dispersive X-ray fluorescence. Analyses of this element by this method have analytical uncertainties of ± 0.01 percent.

Only noncalcareous soil horizons were analyzed for carbon isotopes. Splits of samples from each soil horizon were combusted in an automated Dumas combustion device at ~1,000°C. Resulting combustion gases were passed through a series of chemical scrubbers to remove interfering species, leaving the CO_2 and He carrier. Gases were passed directly into the ionization source and the CO_2 was analyzed on a triple-collecting gas-source mass spectrometer. Results are reported in delta notation relative to the Peedee belemnite (PDB) standard.

Soil Properties Useful in Identifying and Interpreting Buried Soils in Alaskan Loess

In the field, soil morphology (color, texture, structure, consistence, horizonation) is what distinguishes modern soils in Alaska and elsewhere from unaltered sediments or other parent materials. In different climates and biomes of Alaska, soils with well-developed horizons are distinct from unaltered loess, primarily because of different-colored horizons, although there can also be subtle changes in texture, structure, and the abundance of roots. However, recognition of buried soils on the basis of morphology alone may be difficult in fresh exposures of perennially frozen loess.

Organic matter shows consistent depth relations in modern, well-drained soils from a wide variety of climates and biomes, and the trends have been summarized by Jenny (1980). Soil organic matter content is a function of the competing processes of organic-matter production by living organisms and destruction by decomposition. Because production rates are usually highest at the soil surface in well-drained soils, organic matter contents tend to be highest in surface horizons and diminish rapidly with depth: this is therefore a characteristic "depth function" found in many soils. For most well-drained soils, Jenny (1980) suggests that organic-matter-content depth functions are best approximated by exponential curves. In contrast, poorly drained soils frequently have monotonically high organic matter contents to considerable depths, resulting in linear depth functions. Organic matter production rates and decomposition rates in coastal southern Alaska soils are relatively high (Burt and Alexander, 1996). In contrast, both organic matter production rates and decomposition rates are relatively low in tundra and boreal-forest environments in other parts of Alaska (Chapin and others, 1980; Van Cleve and others, 1991). The characteristic exponential depth function of soil organic matter, commonly found in mid-latitude and low-latitude environments, may be a key property used to identify buried soils. However, cryoturbation processes are common in many Alaskan tundra soils and even some boreal-forest soils (Everett and Parkinson, 1977; Bockheim and Tarnocai, 1998; Bockheim and others, 1997, 1998). These mechanisms could produce highly irregular organic matter-depth functions. Nevertheless, the few published studies of organic-matter concentrations in Alaskan soils show depth functions similar to mid-latitude and low-latitude soils (Rieger and others, 1979; Michaelson and others, 1996; Bockheim and others, 1998).

Phosphorus, an important plant nutrient in soils, is derived chiefly from the weathering of apatite [$Ca_5(PO_4)_3(F,Cl,OH)$]. The P released by apatite weathering can enter several pools: small amounts can be leached, some can be taken up in secondary P minerals, some can be taken up by plants and converted to organic forms (many of which are insoluble), some becomes part of the soil labile (exchangeable) pool, and some becomes part of what has been referred to as the occluded (encapsulated in other minerals) pool (Smeck, 1985). Although P is considered to be relatively immobile in soils (Smeck, 1973), soil chronosequence studies have shown that total P decreases in soils over time (Walker and Syers, 1976). For the present study, however, it is important to note that much of the P released by apatite weathering is taken up by plants, converted to organic forms, and returned to soil surface horizons after dieback. Studies of soils in the United States, Canada, and New Zealand show a characteristic total P depth function as follows: (1) enrichments in surface horizons due to biocycling, (2) depletions in subsurface horizons where plants have removed P or it has been leached, and (3) relative enrichments below the zone of depletion where there has been neither plant uptake nor leaching (Smeck, 1973; Runge and others, 1974; Walker and Syers, 1976; Letkeman and others, 1996). Runge and others (1974) successfully used this characteristic total-P depth function to identify buried soils in loess sequences in New Zealand. Studies have shown that, under both tundra (Gersper and others, 1980; Chapin and others, 1980) and boreal forest (Van Cleve and others, 1993), soil P is found largely in organic forms; Bockheim and others (1998) provide some of the few published P depth functions.

Carbon isotopes potentially can be used to infer past vegetation regimes under which buried soils formed. The use of carbon isotopes for paleoecological reconstructions is reviewed by Tieszen (1991). Two types of photosynthetic pathways, C_3 and C_4, fractionate carbon isotopes in distinctly different ways. Plants following the C_3 pathway, which include cool-season grasses, all trees, and most tundra plants, have $\delta^{13}C$ values that range from –22 to –34 ‰ and average about –26 ‰. In contrast, C_4 plants, which include most warm-season grasses, range from –9 to –20 ‰ and average –12 ‰ (O'Leary, 1988). Because Alaska presently has only forest and tundra vegetation (both dominated by C_3 plants), it could be assumed that carbon isotopes would be of little use in interpreting past vegetation types. However, there is a considerable range of variability of carbon isotopic composition even within the realm of C_3 plants. Ugolini and others (1981) reported that tundra litter at one locality in Alaska showed a range of $\delta^{13}C$ values from about –21.8 to –24.0 ‰, whereas boreal forest litter at lower elevations showed a range of values from –25.5 to –27.1 ‰. Although the number of samples (~6 of each biome) was limited in this study, the reported differences in carbon-isotope composition are much greater than typical analytical uncertainty and suggest that the technique has promise for differentiating tundra- and forest-derived buried soils.

Vegetation in the Study Areas

In northern Alaska, the Brooks Range is dominated by alpine tundra, whereas the Arctic Foothills and Arctic Coastal Plain have moist tundra that grades into wet tundra (fig. 3). At most of the localities we studied in the Arctic Foothills and on the Coastal Plain, the vegetation can be characterized as moist tundra. In the interior near Fairbanks, soils studied are found in upland *Picea-Betula-Populus* forest (fig. 4) dominated by white spruce (*Picea glauca*), as well as more poorly drained sites dominated by black spruce (*Picea mariana*). The vegetation above ~900-1,000 m in the Alaska Range consists of both alpine tundra and moist tundra (fig. 5). Moist tundra (low sedge tussock/shrub tundra) is the characteristic vegetation on two soil catenas that were studied. The dominant shrubs are birch (*Betula glandulosa*) and willow (*Salix pulchra*). On the Kenai Peninsula, both interior-type boreal forest and coastal forest are found (fig. 6). However, there is also a transitional community, particularly along a narrow (less than 1 km wide) strip on the west coast near the city of Kenai. This transitional community is dominantly boreal forest, but its understory has some of the elements of coastal forest such as devil's club (*Oplopanax horridum*), *Sorbus* sp., and abundant ferns. Many of the upperstory trees are hybrids of Sitka spruce (*Picea sitchensis*) and white spruce (*Picea glauca*). Most of the Spodosols sampled in the Kenai area are developing under what is presently boreal forest, but two pedons were sampled in the transitional coastal-boreal forest community.

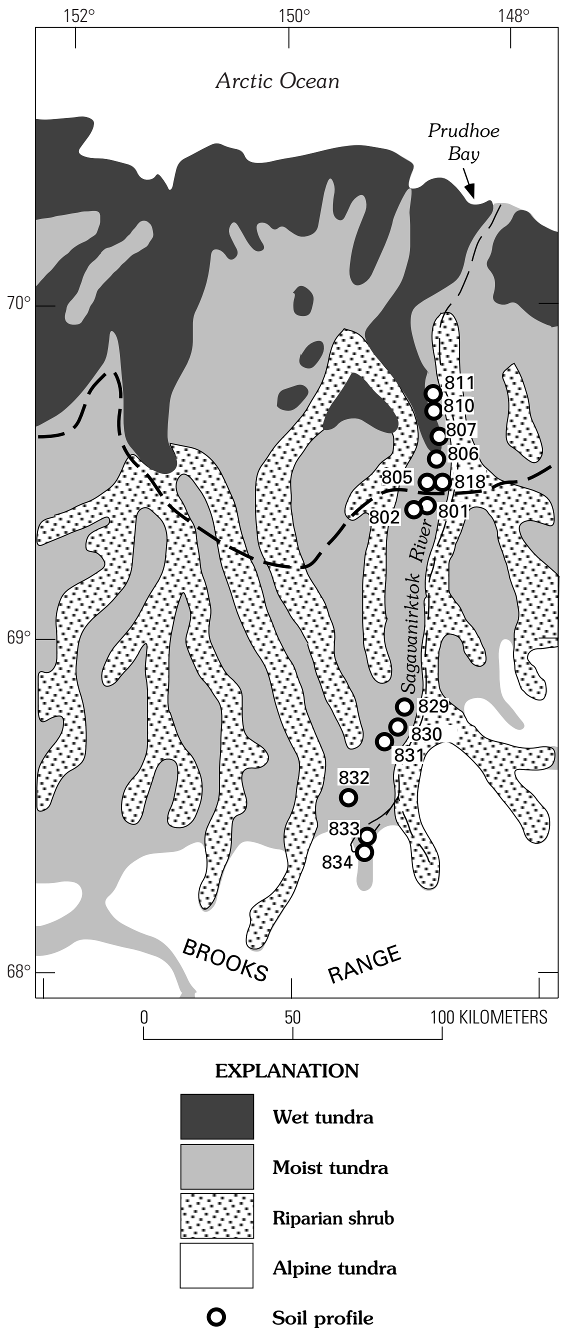

Figure 3. Major ecosystem types in the northern Brooks Range, Arctic Foothills, and Arctic Coastal Plain regions of Alaska, and location of pedons (soil profiles) studied. Dashed line indicates approximate boundary between the Arctic Foothills and the Arctic Coastal Plain. Vegetation distribution from Joint Federal-State Land Use Planning Commission for Alaska (1973).

Geochronological studies indicate that most of the modern soils studied on the Arctic Coastal Plain, in the interior near

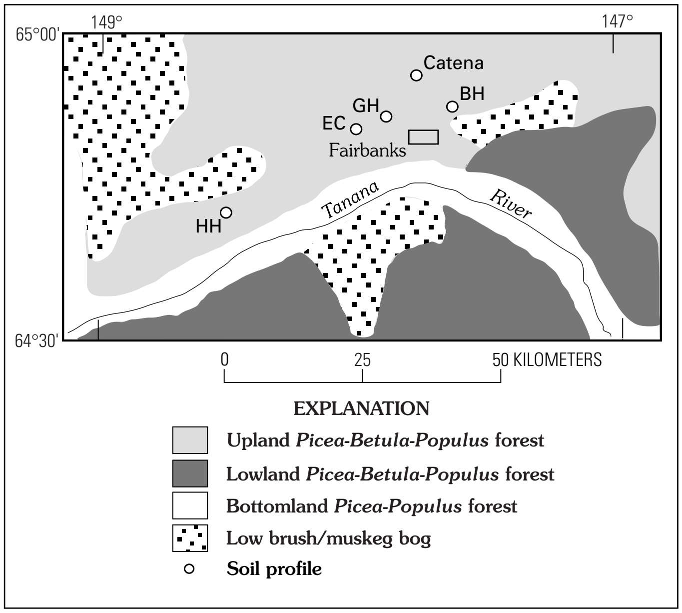

Figure 4. Major ecosystem types in the Fairbanks, Alaska, area and location of soil catena and other pedons studied. HH, Halfway House; EC, Eva Creek; GH, Gold Hill; BH, Birch Hill. Vegetation distribution from Joint Federal-State Land Use Planning Commission for Alaska (1973).

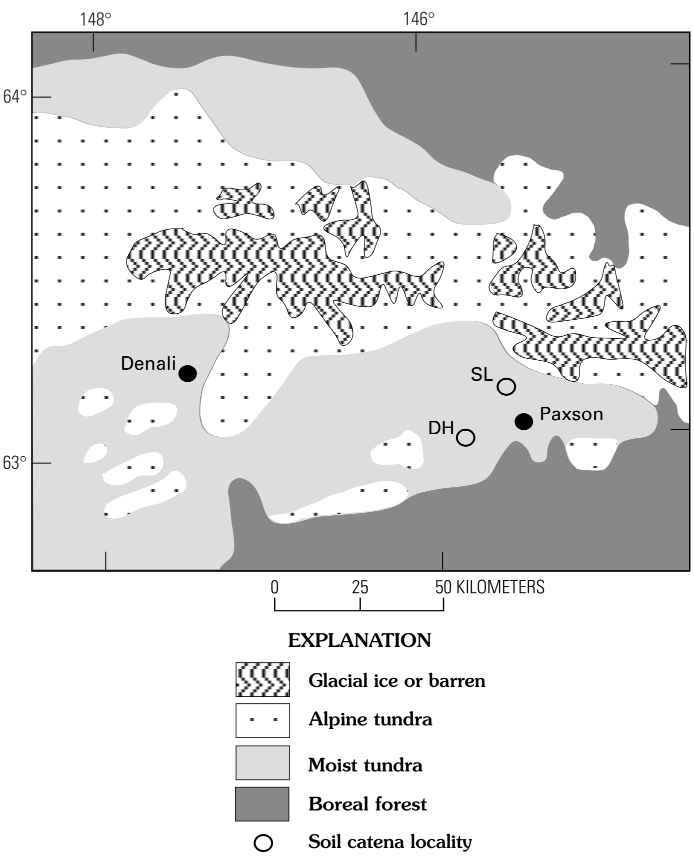

Figure 5. Major ecosystem types in the central part of the Alaska Range and surrounding areas, and location of soil catenas near Summit Lake (SL), and on the Denali Highway (DH) near Ten Mile Lake. Vegetation distribution from Joint Federal-State Land Use Planning Commission for Alaska (1973).

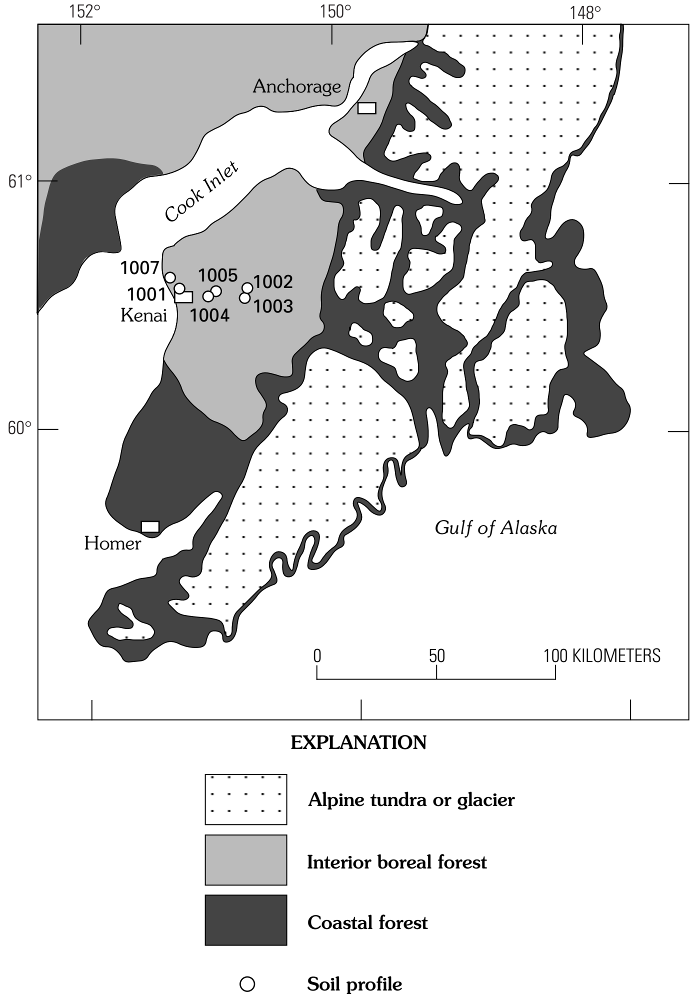

Figure 6. Major ecosystem types in the Kenai Peninsula area and location of pedons studied. Vegetation distribution slightly modified from Joint Federal-State Land Use Planning Commission for Alaska (1973).

Fairbanks, and near Kenai have formed in loess of Holocene or latest Pleistocene age (Everett and Brown, 1982; Brown and Krieg, 1983; Walker and others, 1998; Hamilton and others, 1983, 1988; Begét, 1990; Péwé and others, 1997; Muhs, Ager, and others, 1997; Reger and others, 1996). Pedons in both the Arctic Foothills and the Alaska Range have developed in thin loess over glacial till or outwash that dates to the last glacial period between 24 and 11.5 ka (Coulter and others, 1965; Hamilton, 1978, 1994; Anderson and others, 1994). Pollen studies from all these regions indicate that modern soils have developed primarily under vegetation that is little different from that found in those regions today (Ager, 1983; Ager and Brubaker, 1985; Ager and Sims, 1982; Anderson and Brubaker, 1994; Anderson and others, 1994).

Properties of Soils in the Study Areas

Soil Morphology

Soil morphological properties are what distinguish individual soil horizons from one another. In this paper, terminology for soil horizons and pedon (soil profile)

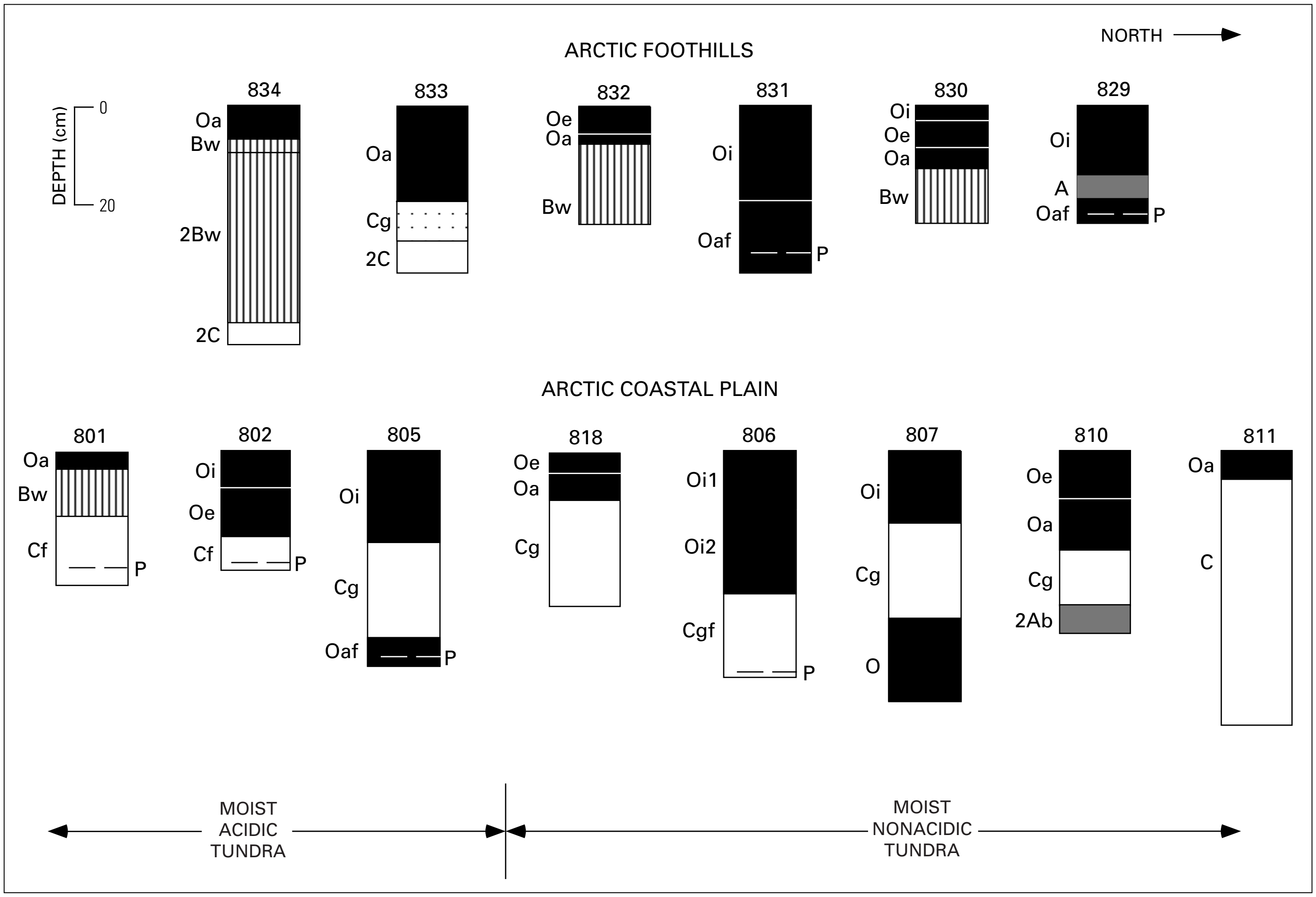

Figure 7. Soil morphology in pedons studied in the Arctic Foothills and Arctic Coastal Plain regions, developed under moist acidic and nonacidic tundra (terminology from Bockheim and others, 1998, and Walker and others, 1998). "P" symbol with dashed line indicates depth of permafrost at time of sampling (July 1996). Pedon numbers keyed to figure 3.

classification follows definitions given by the Soil Survey Staff (1998). On the Arctic Coastal Plain, Gelisols (permafrost-affected soils) occur where permafrost occurs within 100 cm of the surface or within 200 cm of the surface if there is gelic material within 100 cm of the surface (Bockheim and others, 1997; Soil Survey Staff, 1998). Many of the soils studied in this area qualify as Gelisols because permafrost was encountered at depths of ~20-50 cm at the time of sampling in July 1996 (fig. 7). Some of the soils with evidence of cryoturbation are likely classified as Histoturbels (profiles 805 and 806) or Haploturbels (profiles 801 and 802). Other soils without evidence of cryoturbation are most likely Aquorthels (profile 807) or Historthels (profile 810). Soils that do not have permafrost at the requisite depths are Inceptisols.

Most of the Arctic Coastal Plain soils studied have O/Cg profiles, whether developed under moist acidic or nonacidic (terminology from Bockheim and others, 1998 and Walker and others, 1998) tundra (fig. 7). Many soils are poorly drained, and O horizon thicknesses are quite variable. Several pedons studied have buried O or A horizons, similar to other Arctic Coastal Plain soils studied by Michaelson and others (1996) and Bockheim and others (1998). No E horizons were observed in any of these soils, and only one pedon studied (801) had a B horizon.

Soils formed in loess over till or outwash of last glacial age in the Arctic Foothills generally have a greater degree of profile differentiation than soils on the Arctic Coastal Plain (fig. 7). At least two of these pedons (829 and 831) are Gelisols, but others with deeper permafrost are likely Dystrocryepts (830 and 834). No E horizons or Bt horizons were observed in Arctic Foothills soils.

In the Alaska Range, soils developed in loess over till show variability as a function of slope position (fig. 8). Some of these soils may qualify as Gelisols, but the depth to permafrost is not known in most of our study sites. Relatively well drained soils on crest and shoulder positions and even upper footslope positions have thin O horizons and brown or yellowish-brown Bw horizons. At poorly drained lower footslope positions, soils have minimally developed grayish-brown B horizons or lack B horizons altogether. As in soils in the Coastal Plain and the Arctic Foothills, E horizons were not observed in tundra soils of the Alaska Range, except possibly in the ridge-crest position soil in the Summit Lake catena (fig. 8). This thin, light-colored horizon may be unaltered loess, but given a possibility of slightly higher treeline in this area in the early Holocene (Anderson and others, 1994), we cannot exclude the possibility that it is an E horizon that developed under a former forest cover.

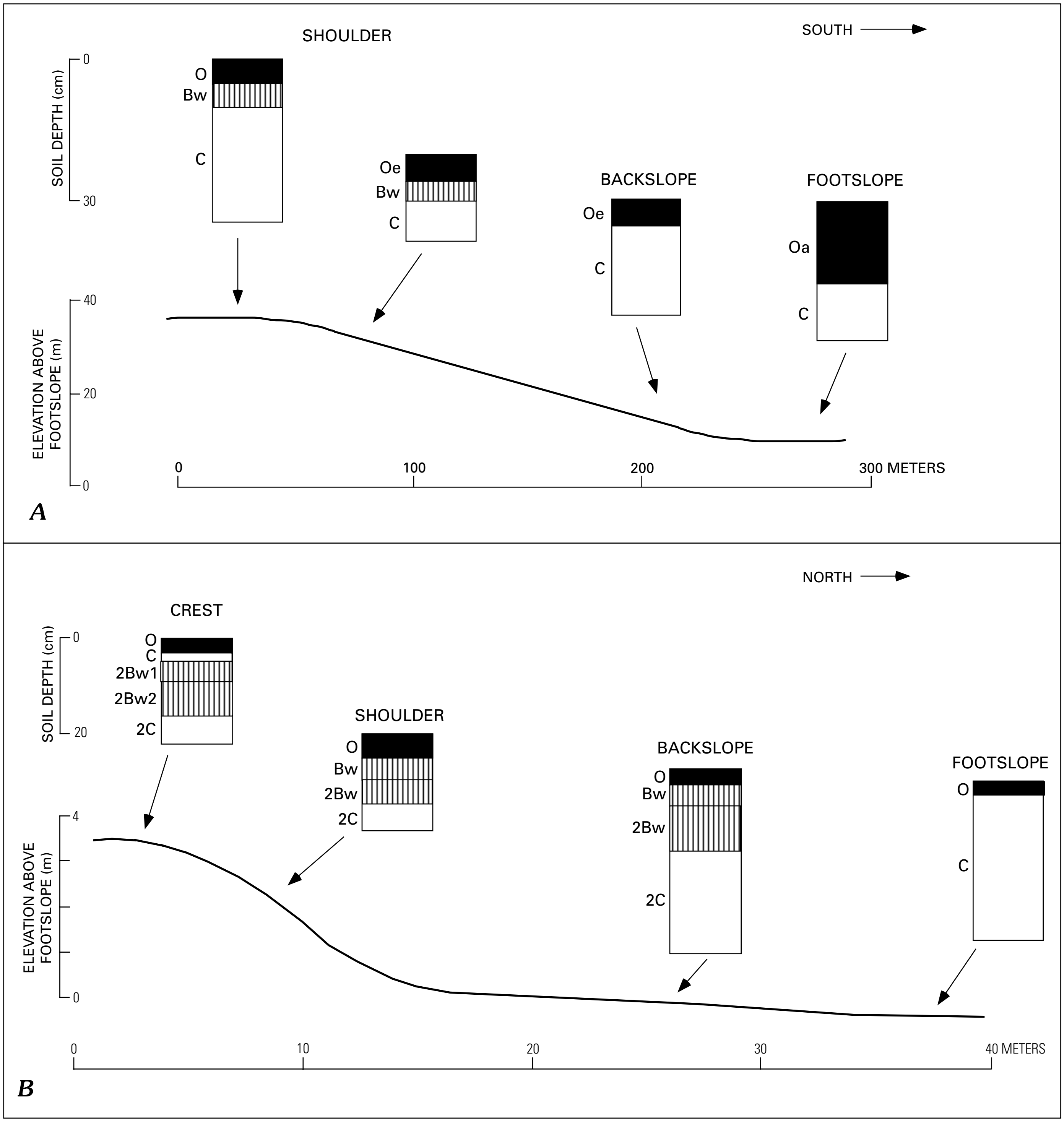

Figure 8. Soil morphology along catenas on *A*, the Denali Highway near Ten Mile Lake and *B*, near Summit Lake in the Alaska Range. Locations on figure 5.

Inceptisols and Gelisols developed under interior boreal forest near Fairbanks show distinct horizonation that varies with slope position (fig. 9). The greatest degree of profile differentiation is at the shoulder and upper backslope positions, where soils have A/E/Bw/Bt/C or A/E/Bw/C profiles. All soils at lower backslope or higher positions are Typic Dystrocryepts that lack shallow permafrost. In all soils downslope of the lower backslope position, E horizons are absent. At upper footslope positions, soils are Aquic Dystrocryepts with O/Bw/C profiles, and poorly drained soils at the lower footslope position are Aquic Historthels with Oi/Cg profiles. Permafrost was encountered at depths of 70–90 cm in lower footslope soils at the time of sampling (September 1996).

Spodosols (Haplocryods) developed under coastal-boreal forest or boreal forest near Kenai (Rieger and others, 1979) have very distinct horizons (fig. 10). O horizons are very dark (10YR 2/1 or 2/2) even when dry. All Spodosols studied have E horizons that are more distinct than under interior boreal forest and

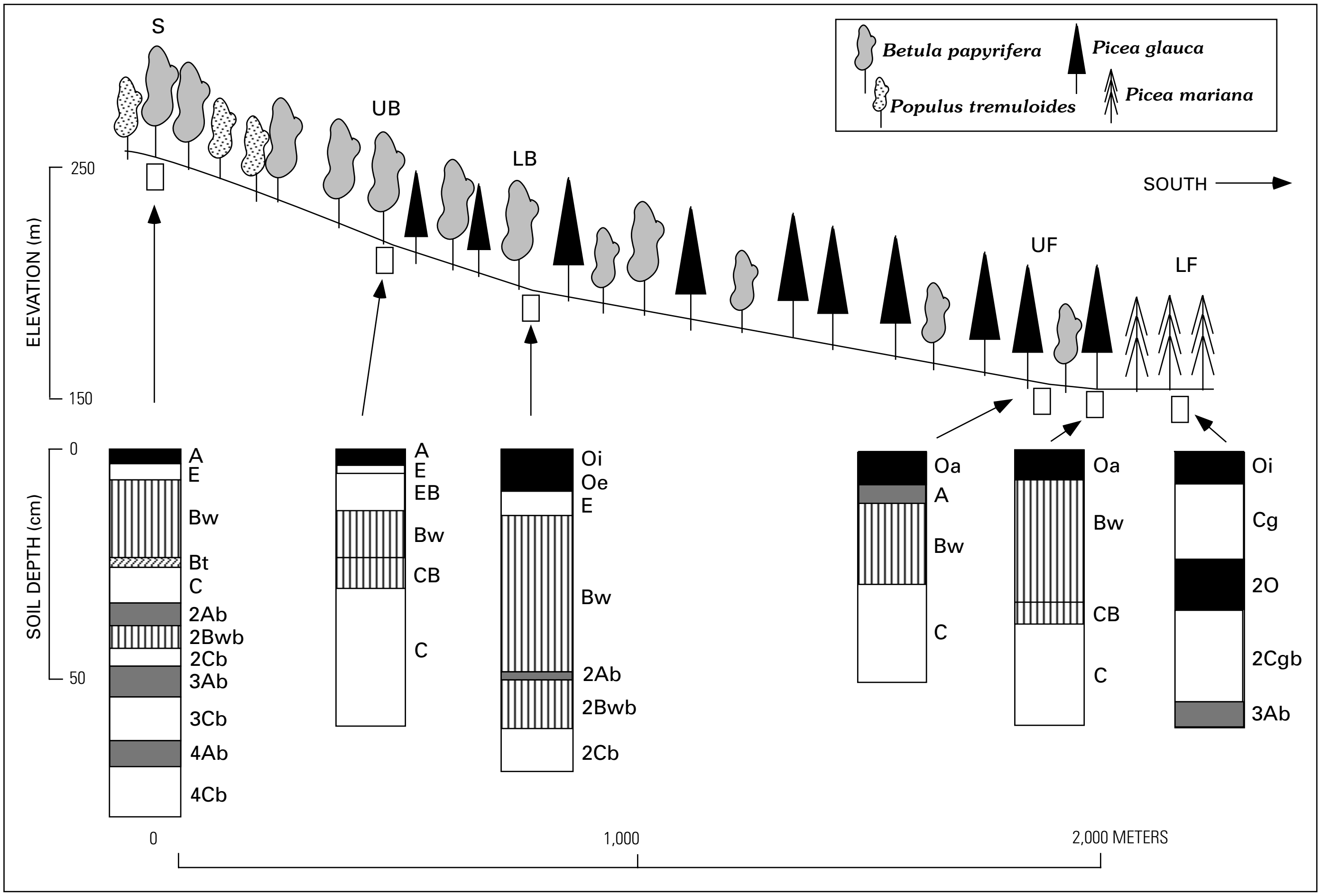

Figure 9. Soil morphology along a catena north of Fairbanks and dominant boreal forest taxa. Slope positions: S, shoulder; UB, upper backslope; LB, lower backslope; UF, upper footslope; LF, lower footslope. Location on figure 4.

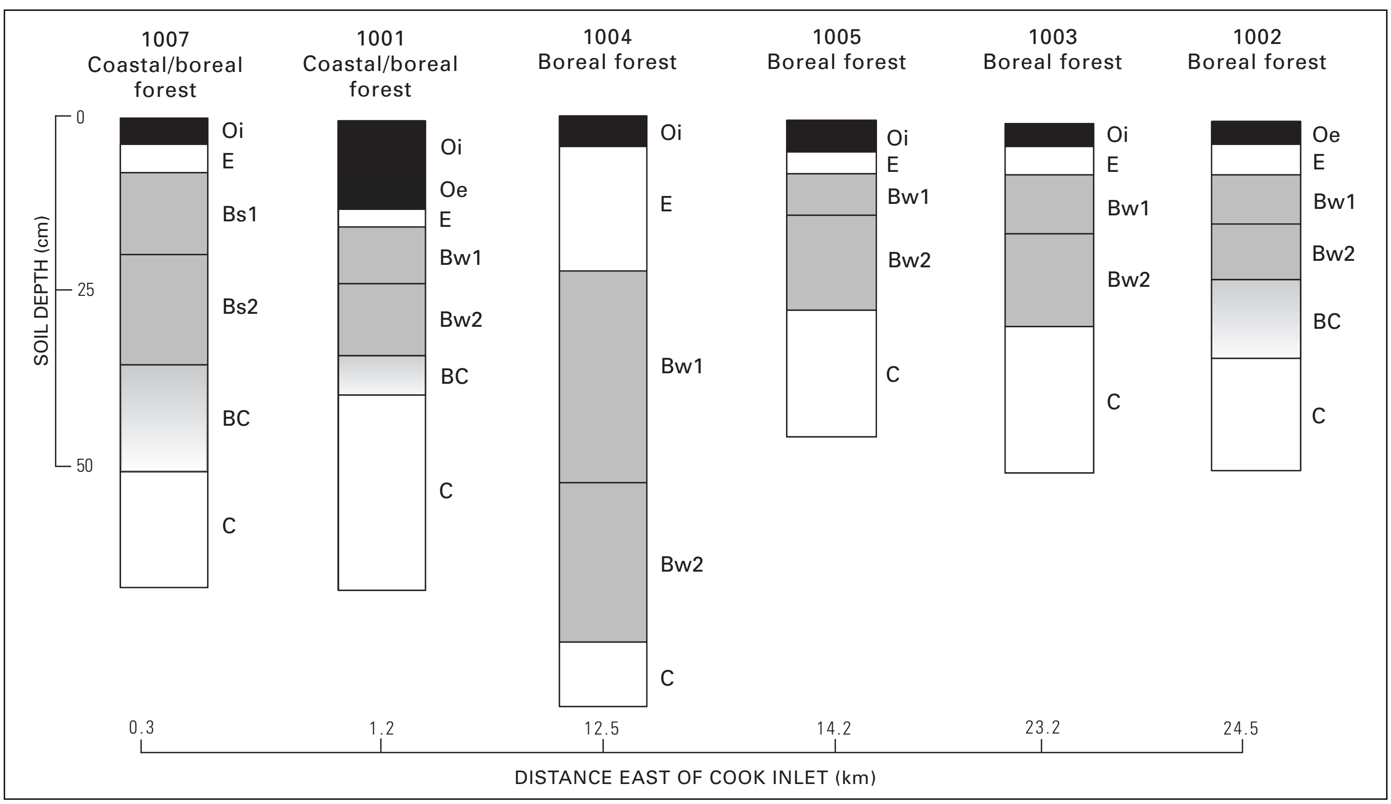

Figure 10. Soil morphology in pedons studied in a transect from Cook Inlet to ~24 km inland on the Kenai Peninsula. Profile numbers keyed to figure 6.

are underlain by Bw horizons. Near the coast, under transitional coastal-boreal forest, Spodosols have Bs horizons that are much higher in total Fe_2O_3 content than the parent loess or overlying E horizon (Muhs and others, in press). Although all Spodosols studied have Bw or Bs horizons, no Bt horizons were observed.

Studies of modern, well-drained Alaskan soils indicate that morphological characteristics, particularly horizon sequences with their distinctive colors, can be used to identify buried soils in unfrozen, dry exposures of loess. Unaltered loess in Alaska typically has light brownish gray (2.5Y 6/2, dry), light yellowish brown (2.5Y 6/3, dry) or grayish-brown (2.5Y 5/2, dry) colors. Soil O or A horizons generally have 10YR or 7.5YR hues and much lower values and chromas (usually 1 or 2) that distinguish them from unaltered loess. Well-drained soils with Bw and Bs horizons generally have redder hues (10YR or 7.5YR) and higher chromas (3–6) than unaltered loess. In forests, soil E horizons are very distinctive features (10YR or 7.5YR 6/1 or 6/2 colors) in Spodosols and well-drained Inceptisols that contrast with dark, overlying O horizons—these E horizons ought to be identifiable in loess-derived buried soils.

Organic-rich alluvial, colluvial, or lacustrine deposits that could be mistaken for buried soils in a stratigraphic section may superficially resemble O or A horizons of soils. However, these deposits are unlikely to have horizon sequences that would be mistaken for a soil O/E/Bw/C horizon sequence. On the other hand, buried soils with only O/Cg profiles that developed under poorly drained conditions could be mistakenly identified as organic-rich sediments, and, in such cases, other criteria (discussed below) need to be included for positive identification as a buried soil.

It is more difficult to interpret past vegetation type on the basis of buried soil morphology. Dark soil O horizons, frequently of similar thicknesses, are present under tundra, boreal forest, and coastal-boreal forest, so surface horizons say little about vegetation type. Tundra soils lack E horizons, whereas these horizons are found in all Spodosols and well-drained Inceptisols that developed under forest. However, boreal forest soils that developed in poorly drained landscape positions lack E horizons. Therefore, while the presence of an E horizon in a buried soil implies a past forest vegetation, the lack of an E horizon in a buried soil does not necessarily imply tundra vegetation, unless there is certainty that the former land surface was well drained. In well-drained landscape positions, both tundra and forest soils can have Bw horizons; thus, Bw horizons cannot be used to infer past vegetation type.

Organic Matter

Organic matter distributions in both acidic and nonacidic tundra soils from the Arctic Coastal Plain and Arctic Foothills show depth functions typical of well-drained mid-latitude and low-latitude soils in some cases, but not in others (fig. 11). Organic matter contents are extremely high (>70 percent in some soils) in surface O horizons, and many of these are thick enough to qualify as histic epipedons. Concentrations of organic matter diminish rapidly as a function of depth in some soils (e.g., profiles 801, 810, 830, 834), but less so in others (profiles 805, 806, 807). The soils that do not show rapid organic matter decreases with depth are those that are very poorly drained and have relatively thick Oi horizons. Some of the latter soils were saturated and had standing water at the time of sampling.

In modern boreal forest soils in the Fairbanks area, O horizons are generally thinner than those developed under tundra and, in some cases, notably at better drained positions, have organic matter contents that are lower than those in tundra soils (fig. 12). Organic matter depth functions in boreal forest soils show the characteristic exponential decreases as described by Jenny (1980). A buried O horizon, identifiable by its very dark grayish brown (10YR 3/2, moist) or black (10YR 2/1, moist) color in the lower footslope position is also clearly marked by its higher organic matter content (fig. 12). However, thin buried soils that were identified in the field at both backslope and shoulder positions by their darker colors do not have increases in organic matter. Therefore, at well-drained slope positions, organic matter content sometimes may be a poor indicator of the presence of a buried soil.

Modern tundra soils at the Summit Lake catena in the Alaska Range also show relatively rapid declines in organic matter as a function of depth (fig. 13). As is the case with boreal forest soils in the Fairbanks area, organic matter contents are highest at lower slope positions (backslope and footslope), but the forms of depth functions at all slope positions are similar.

Spodosols under either boreal forest or transitional coastal-boreal forest have O horizons with organic matter contents of 30–70 percent that decrease very rapidly with depth (fig. 14). Two Spodosols (profiles 1001 and 1007) that developed under transitional coastal-boreal forest and one that developed under boreal forest (profile 1005) have very distinct decreases in organic matter content in their E horizons, but they show slight enrichments in the upper Bs or Bw horizons. Such translocation of organic matter is characteristic of some Spodosols (Rieger, 1983).

Overall, the depth functions observed here suggest that organic matter concentration is a valuable property for identifying buried soils in loess sections. All forest soils and most tundra soils show decreases in organic matter content as a function of depth, very similar in form to the characteristic organic matter depth functions that have been described for well-drained mid-latitude and low-latitude soils. Organic-rich colluvial, alluvial, or lacustrine deposits found in a stratigraphic section are more likely to show irregular changes in organic matter content as a function of depth. However, some buried boreal forest soils that were identified on the basis of morphology could not be detected by organic matter trends, and some poorly drained tundra soils do not show rapid decreases in organic matter with depth. Finally, because organic matter in most well-drained Alaskan soils decreases rapidly with depth, buried soils whose O or A horizons have been partially or wholly eroded might not be detected in a stratigraphic section. Therefore, organic matter content should be used in conjunction with other properties to identify buried soils.

Phosphorus

Phosphorus distributions in tundra soils of the Arctic Coastal Plain and Arctic Foothills show trends that generally

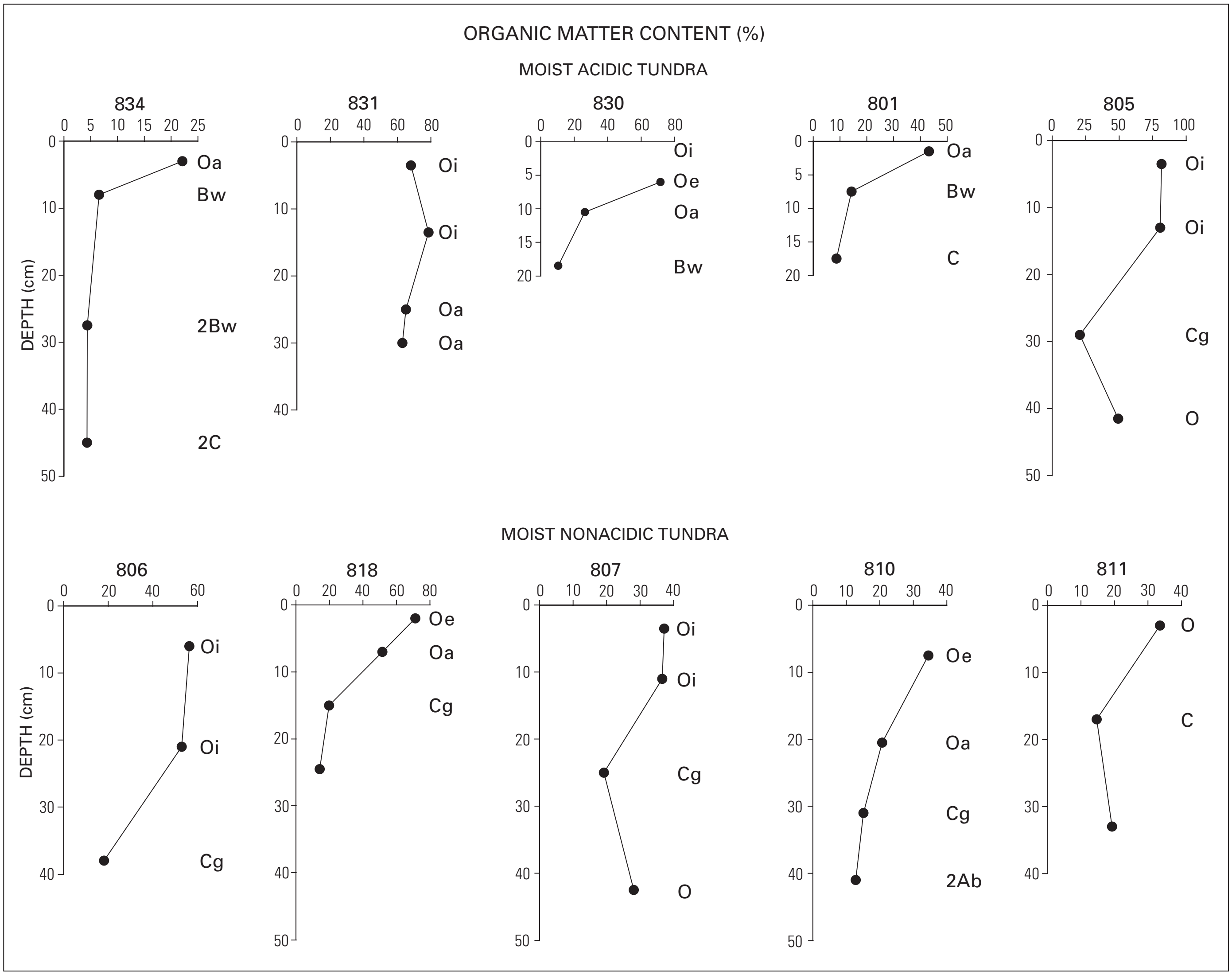

Figure 11. Organic matter content vs. depth in Gelisols and Inceptisols from the Arctic Foothills and Arctic Coastal Plain, developed under moist acidic and nonacidic tundra. Pedon numbers keyed to figures 3 and 7.

parallel organic matter distributions (fig. 15). The same is true for tundra soils in the Summit Lake catena of the Alaska Range, with the exception of the O horizon at the backslope position (fig. 13). Concentrations of P are highest in surface O horizons and diminish rapidly with depth, supporting earlier interpretations (Gersper and others, 1980; Chapin and others, 1980) that P in Alaskan tundra soils is largely in organic forms. Buried O or A horizons in these soils show slight P enrichments compared to overlying C horizons. None of the tundra soils examined show the "high-low-high" surface-to-subsurface depth functions of P concentrations described by Runge and others (1974) and other workers studying mid-latitude soils.

Modern boreal forest soils from the Fairbanks area, on the other hand, show P distributions that are similar to those reported for mid-latitude soils in North America and New Zealand (fig. 12). The highest P values are found in the O or A horizons. The lowest values are found just below the surface horizons and values increase again slightly in the lower B or upper C horizons. The high values associated with the O or A horizons (including the buried O horizon in the lower footslope soil) support the suggestion of Van Cleve and others (1993) that P in boreal forest soils, like tundra soils, is found mainly in organic forms. Spodosols from the Kenai area also show distinct enrichments in P in surface horizons (fig. 14). Most pedons studied show relatively rapid decreases as a function of depth; only two profiles (1007 and 1005) have E horizons with lower values than all other subsurface horizons. In the case of profiles 1007 and 1005 (and perhaps in the lower backslope soil near Fairbanks), it appears that P has been translocated downward in the profile, perhaps in association with organic matter.

Our studies indicate that P concentrations can be a valuable tool in identifying buried soils in Alaska. Although only the interior boreal forest soils show similarities to the "high-low-high" depth function reported for mid-latitude soils, all pedons, whether developed under tundra or forest, show greatest P concentrations in O or A horizons. The importance of these findings to stratigraphy is that P enrichments (succeeded at depth by depletions) should mark surface horizons of buried soils. However, the same problem associated with organic matter depth functions applies to P trends: buried soils with eroded O or A horizons may go undetected.

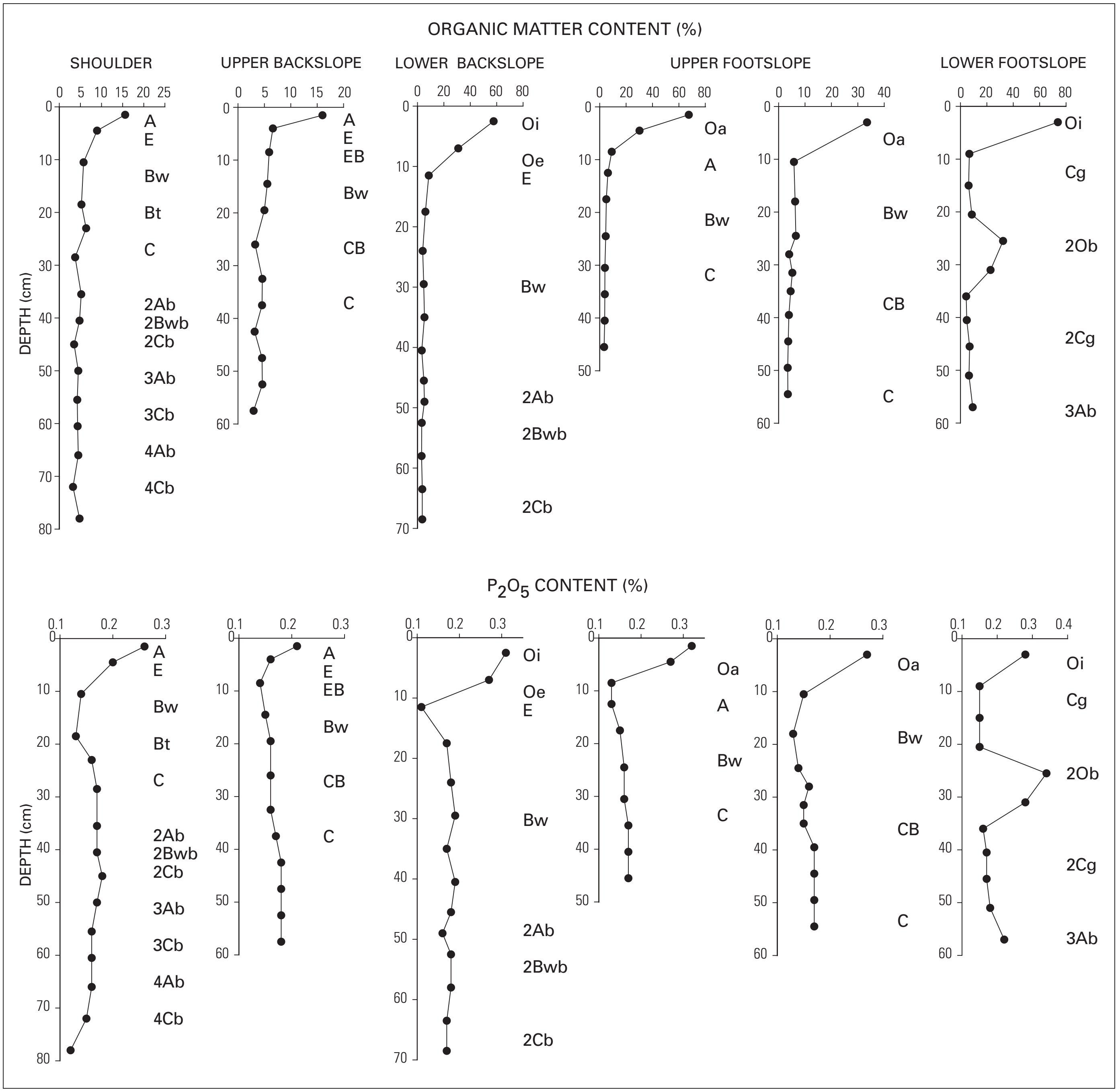

Figure 12. Organic matter and P_2O_5 content vs. depth in Inceptisols along a catena in the Fairbanks area. Slope Positions keyed to figure 9.

Carbon Isotopes

Because buried soil morphology alone often does not permit identification of the dominant vegetation at the time of soil formation, the use of carbon isotopes in vegetation reconstruction from buried soils is tested here. Surface or near-surface O, A, and E horizons should have $\delta^{13}C$ values that are most closely related to modern vegetation. Values from C horizons can reflect the isotopic composition of detrital or in situ vegetation that was in the region during loess deposition, rather than vegetation present during the period of pedogenesis.

Tundra soils from the Arctic Coastal Plain and the Arctic Foothills give $\delta^{13}C$ values ranging from –25.5 to –27.1 ‰ (fig. 16). The uppermost parts of the O horizons have a rather narrow range of values from –26.0 to –26.9 ‰. For some pedons, herbaceous tundra vegetation itself, primarily grasses and sedges, was also analyzed. These materials gave $\delta^{13}C$ values of –25.5 ‰ (profiles 801 and 802) and –26.5 ‰ (profiles 805 and 806) and are similar to $\delta^{13}C$ values of the subjacent horizons. There are no apparent depth-related trends in carbon isotopic composition.

Tundra soils in the Alaska Range have $\delta^{13}C$ values that are very similar to tundra soils in the Arctic Foothills and the Arctic

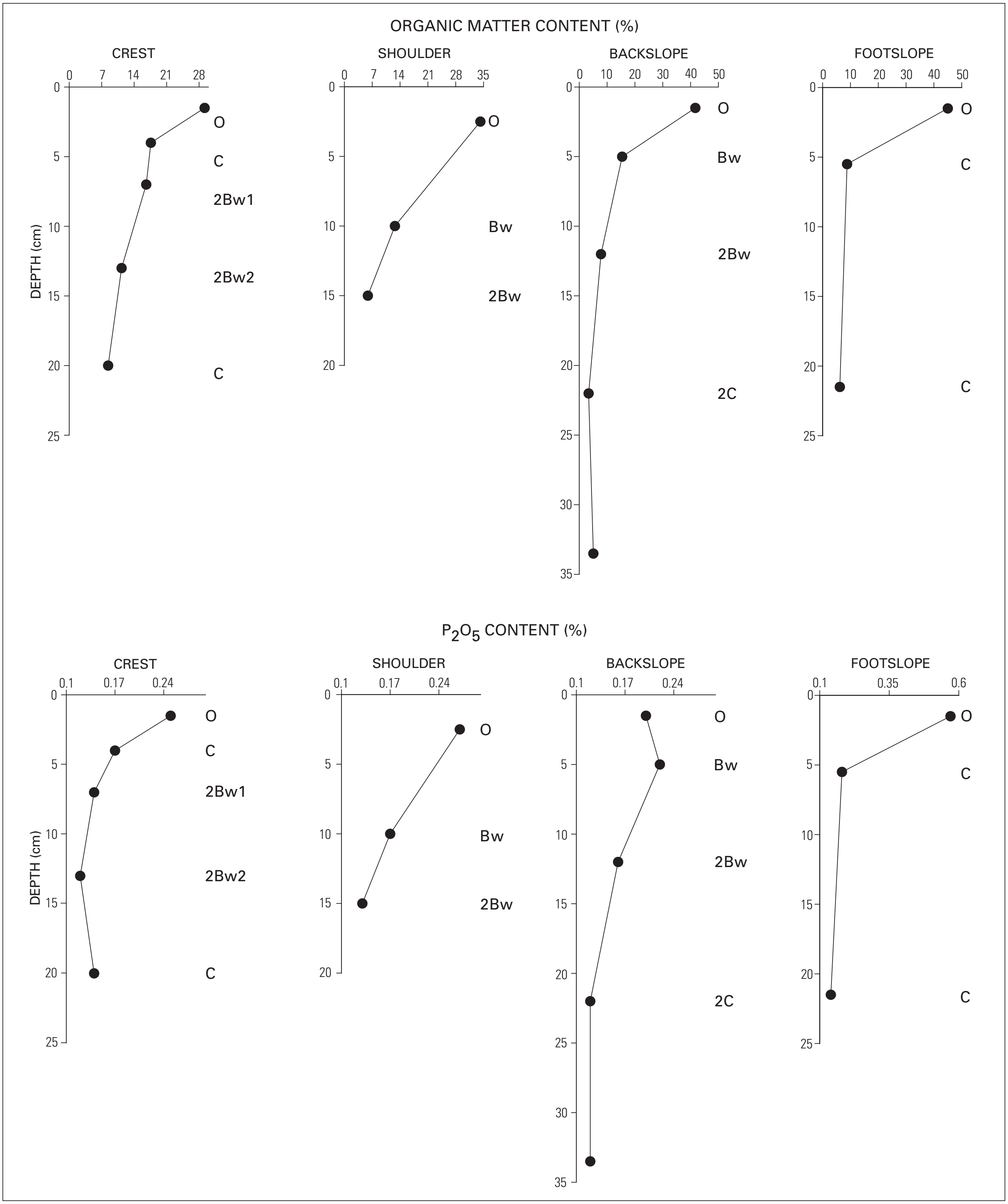

Figure 13. Organic matter and P_2O_5 content vs. depth in soils in the Summit Lake catena in the Alaska Range. Slope Positions keyed to figure 8.

Coastal Plain (fig. 17). The overall range of $\delta^{13}C$ values at both the Summit Lake and Denali Highway catenas is greater than in the Arctic Foothills and Coastal Plain soils (–23.9 to –27.8 ‰), but values are judged not to be significantly different. There are no consistent depth-related trends in carbon isotopic values in the Alaska Range catenas. Although pollen evidence suggests the possibility that some of these soils may have developed partly under a brief period of open boreal forest or forest-tundra

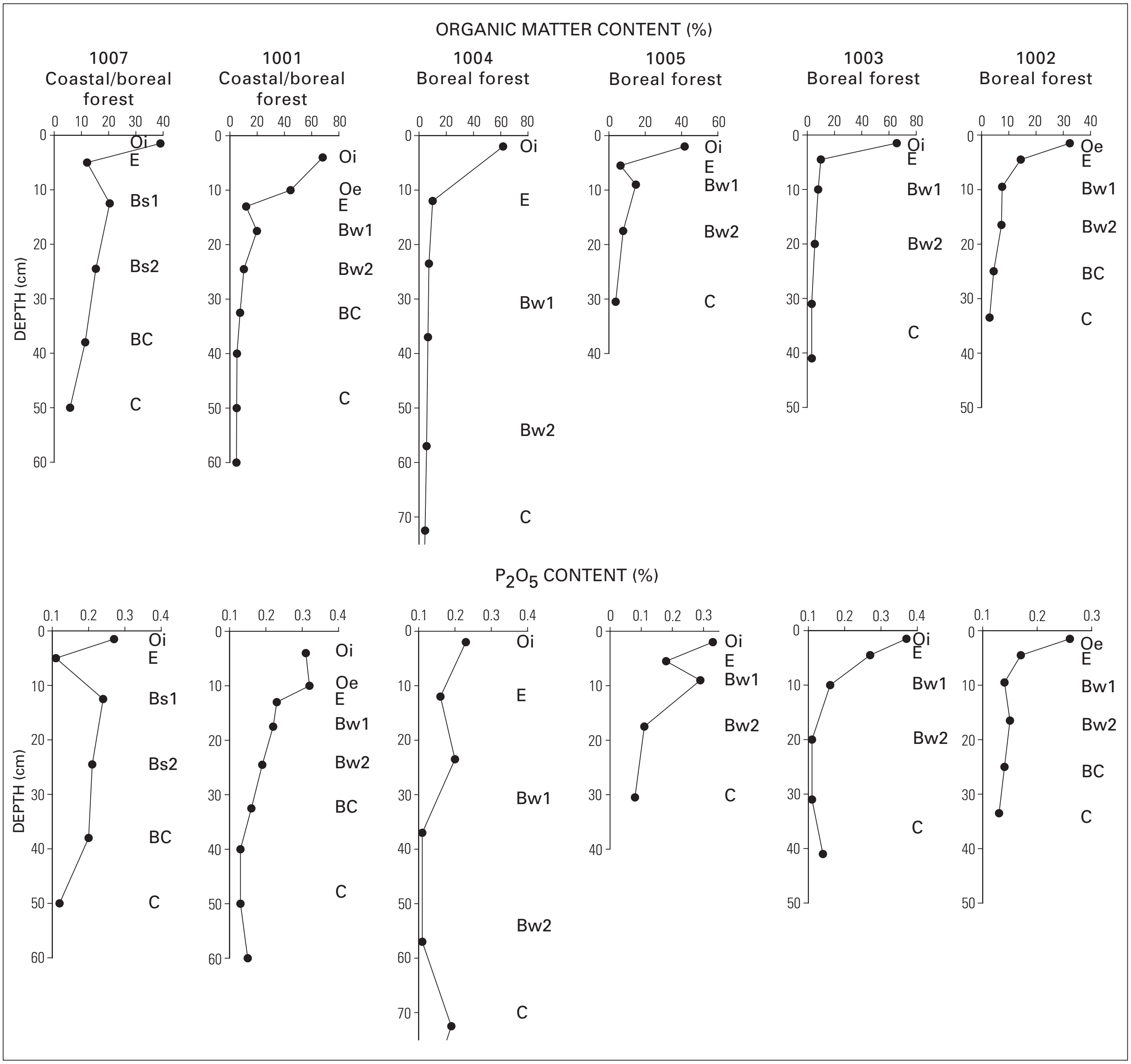

Figure 14. Organic matter and P_2O_5 content vs. depth in Spodosols in the Kenai Peninsula area. Pedon numbers as in figures 6 and 10.

(Anderson and others, 1994), the $\delta^{13}C$ values of the uppermost horizons almost certainly reflect the dominant shrub tundra vegetation that is growing in the area now.

In the Fairbanks area, five modern soils in an upland spruce-birch forest setting were analyzed (localities shown in fig. 4). Carbon isotope values for these soils range from about –25.4 to –27.7 ‰ (fig. 18) and differ little from the values within the uppermost C horizons. As is the case with the tundra soils in the Alaska Range, there are no obvious depth-related changes in carbon isotopic composition.

Spodosols from the Kenai area show the greatest range of carbon isotope values of any of the soils studied. Values range from –24.0 to –29.3 ‰, and O horizons alone span most of this range (fig. 19). Unlike soils from the other regions, Spodosols show a possible trend of isotopically heavier values with greater depth in the profile.

Comparison of the four study areas indicates that carbon isotopes cannot be used to distinguish forest from tundra vegetation in Alaskan buried soils. Uppermost O horizons of tundra soils from all areas studied range from –24.8 to –27.4 ‰, similar to the range of –25.5 to –28.6 ‰ for the uppermost O horizons from all forest soils studied. The results presented here differ from those of Ugolini and others (1981), who found that tundra litter (O horizons) had values of –21.8 to –24.0 ‰ and boreal forest litter had values of –25.5 to –27.1 ‰. Carbon isotope values for both tundra and forest soils in the present study are closest to Ugolini and others' (1981) boreal forest values. However, some Inceptisols and Spodosols that developed under forest

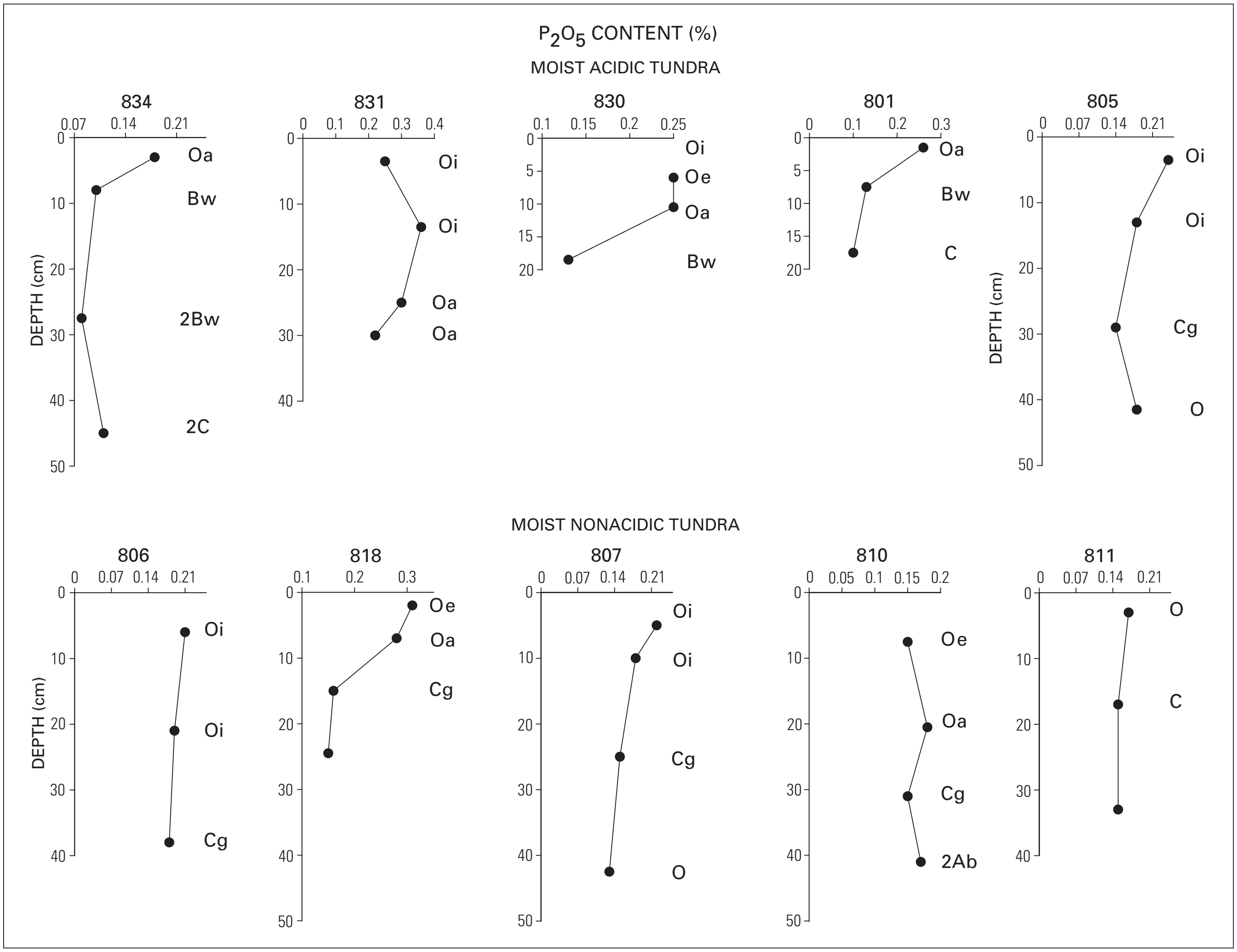

Figure 15. P_2O_5 content vs. depth in Gelisols and Inceptisols from the Arctic Foothills and Arctic Coastal Plain, developed under moist acidic and nonacidic tundra. Pedon numbers as in figures 3 and 7.

vegetation have values greater than –25.0 ‰, which falls within the range of values that Ugolini and others (1981) reported for tundra. None of tundra soils in the present study have $\delta^{13}C$ values as heavy as the tundra values reported in the earlier study, and most have values within the range of what Ugolini and others report for boreal forest. However, the study by Ugolini and others (1981) was based on a very limited number of samples in one local area. The larger sample size reported here shows that the carbon isotope values of soils developed under the two ecosystems are judged not to be significantly different.

Conclusions

Our studies indicate that soil morphology, organic matter concentrations, and P concentrations can be useful diagnostic tools for identifying buried soils. Soil-horizon colors are significantly different from loess parent material, and soil-horizon sequences (e.g., O/Bw/C and O/E/Bw/C profiles) are important clues for identifying buried soils. Nevertheless, soils with only O/C profiles could be misidentified as organic-rich alluvial, colluvial, or lacustrine deposits, and other criteria should be used in conjunction with field data in studying such soils. Furthermore, soil morphology may not be particularly useful for identifying the type of vegetation under which a buried soil may have formed. Although certain soil horizons, such as E and Bs horizons, seem to form only under forest, not all modern forest soils have such horizons. In addition, both well-drained tundra and boreal forest soils can have similar O/Bw/C horizon sequences.

Organic matter concentrations in well-drained Alaskan tundra and forest soils show the rapid decline with depth that is typical of many mid-latitude and low-latitude soils, and this decline is therefore a potentially useful characteristic in identifying buried soils. However, some poorly drained tundra soils show only gradual declines in organic-matter concentrations with depth and thus may be difficult to distinguish from organic-rich sediments in a loess section. Furthermore, because of the rapid decline of organic-matter content with depth, partly eroded buried soils may go undetected in stratigraphic sections. We conclude, therefore, that organic matter trends are a helpful but insufficient tool for identifying buried soils in loess sections.

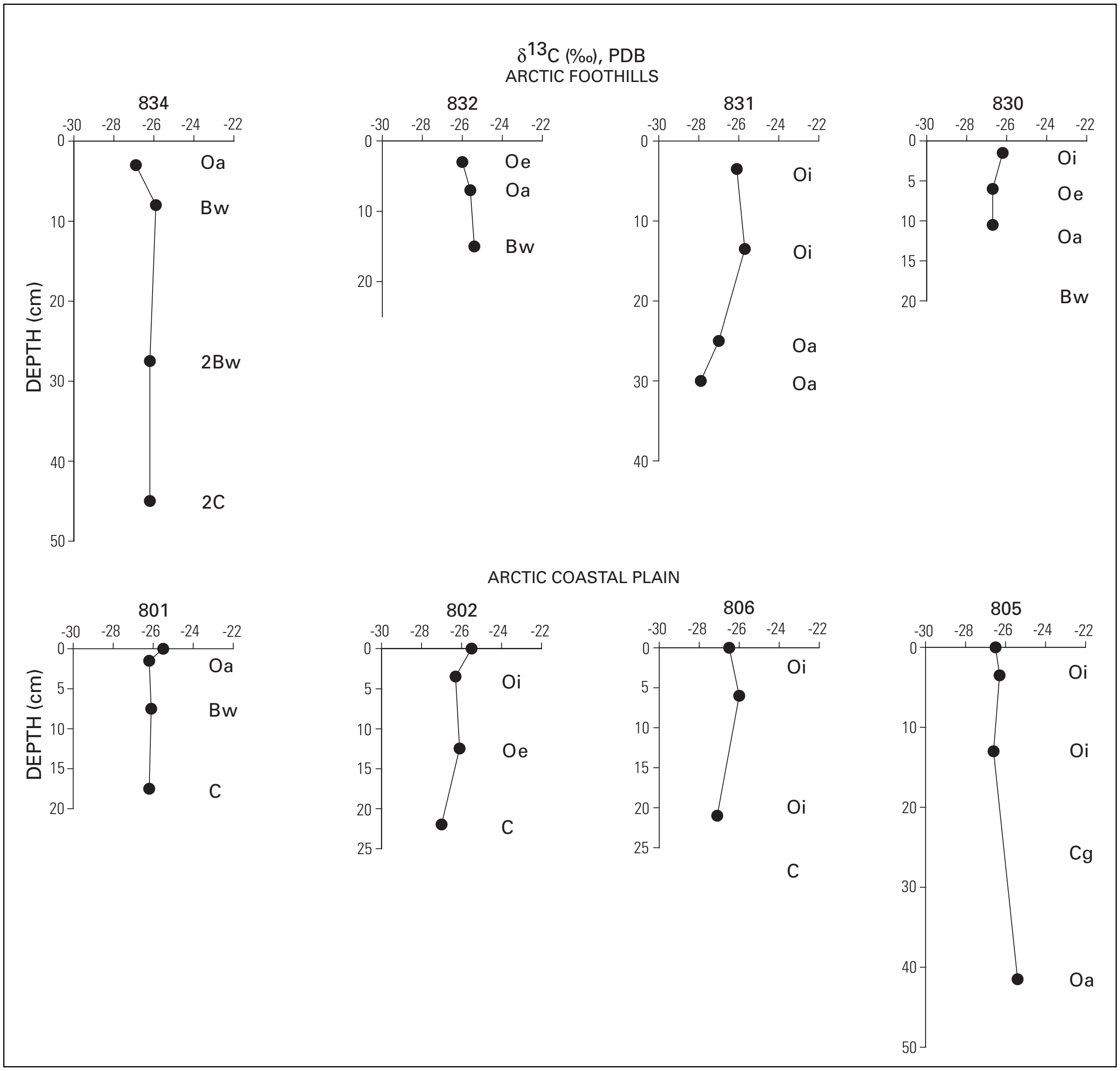

Figure 16. Carbon isotopic composition of soil organic matter vs. depth in Gelisols and Inceptisols from the Arctic Foothills and Arctic Coastal Plain, developed under moist acidic and nonacidic tundra. Pedon numbers as in figures 3 and 7.

In general, concentrations of P follow the organic matter trends, supporting earlier conclusions that P in Alaskan soils is largely in organic forms. In interior Alaska, boreal forest soils show P depth distributions similar to those of mid-latitude soils, with distinct enrichments in surface horizons, depletions immediately below the surface, and slight enrichments again at depth. In contrast, tundra soils and Spodosols show P enrichments in O horizons, but tend to decline consistently with depth. A few poorly drained tundra soils that were studied show no particularly diagnostic depth functions. We conclude that P depth functions are useful tools for identifying buried soils, particularly when used in conjunction with soil morphology and organic matter–depth functions.

Carbon isotope values do not support findings from a previous study that tundra and boreal forest soils can be differentiated using this property. Results from the present study show that carbon isotope values of soil organic matter in tundra and forest soils are not significantly different. Other soil properties need to be investigated to see if they are diagnostic of pedogenesis under forest vs. tundra. Shroba and Birkeland (1983), studying soils developed in late Wisconsin till in the Rocky Mountains, found that vermiculite and other pedogenic clay minerals form under spruce-fir forest, but not under alpine tundra at higher elevations. Bozarth (1993) has shown that boreal forest vegetation has a distinctive assemblage of opal phytoliths that may differentiate it from tundra vegetation. Grass and sedge-dominated tundra may be distinguished from boreal forest using lignin phenols and if these components of organic matter are preserved in soils, may be a fruitful approach for the interpretation of buried soils (Ugolini and others, 1981; Orem and others, 1993, 1997). Finally, fractions of soil organic acids may have a distinctive signature that would allow differentiation of forest from tundra (Ping and others, 1997).

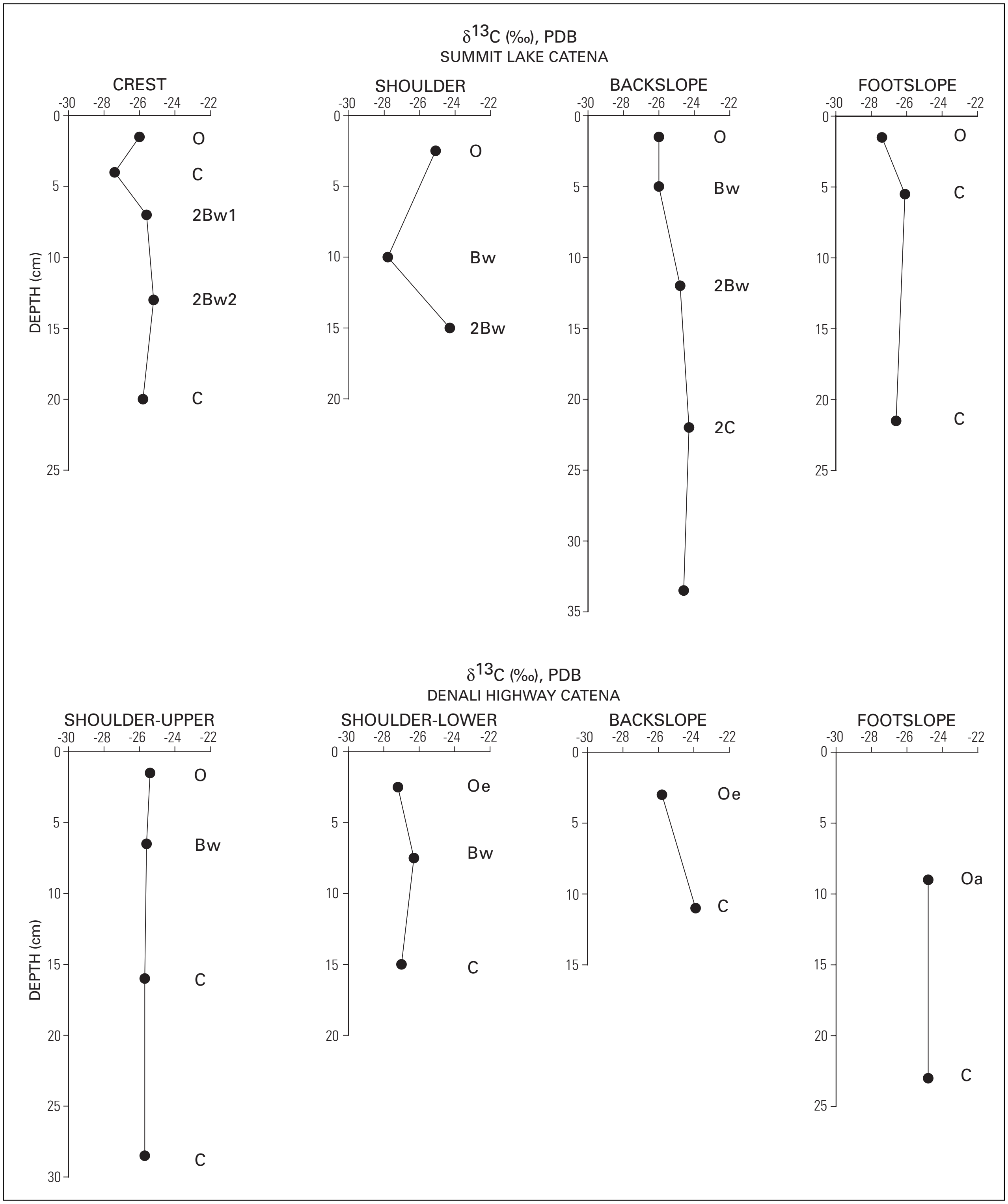

Figure 17. Carbon isotopic composition of soil organic matter vs. depth in tundra soils from catenas in the Alaska Range. Slope positions as in figure 8.

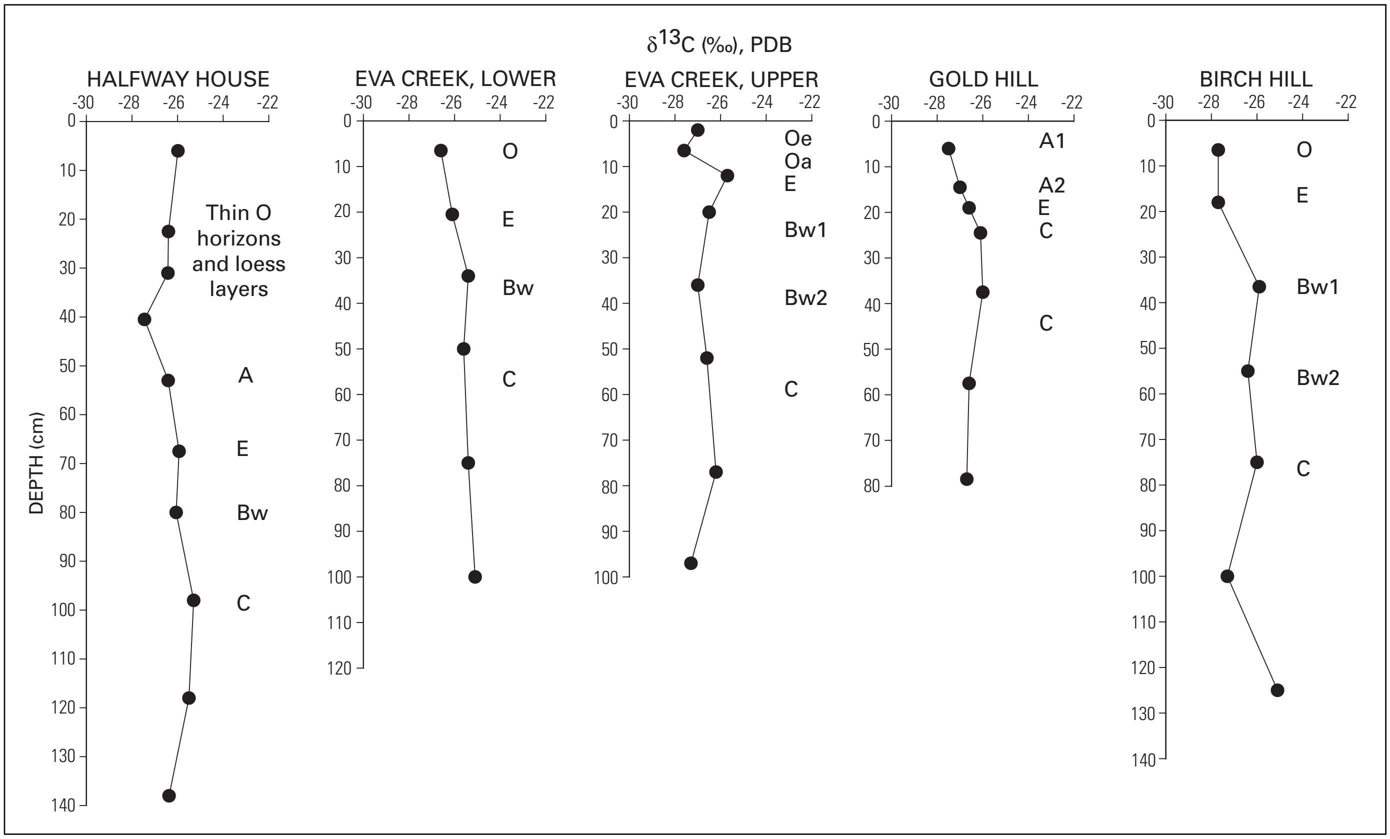

Figure 18. Carbon isotopic composition of soil organic matter vs. depth in five upland boreal forest Inceptisols in the Fairbanks area (localities shown in figure 4).

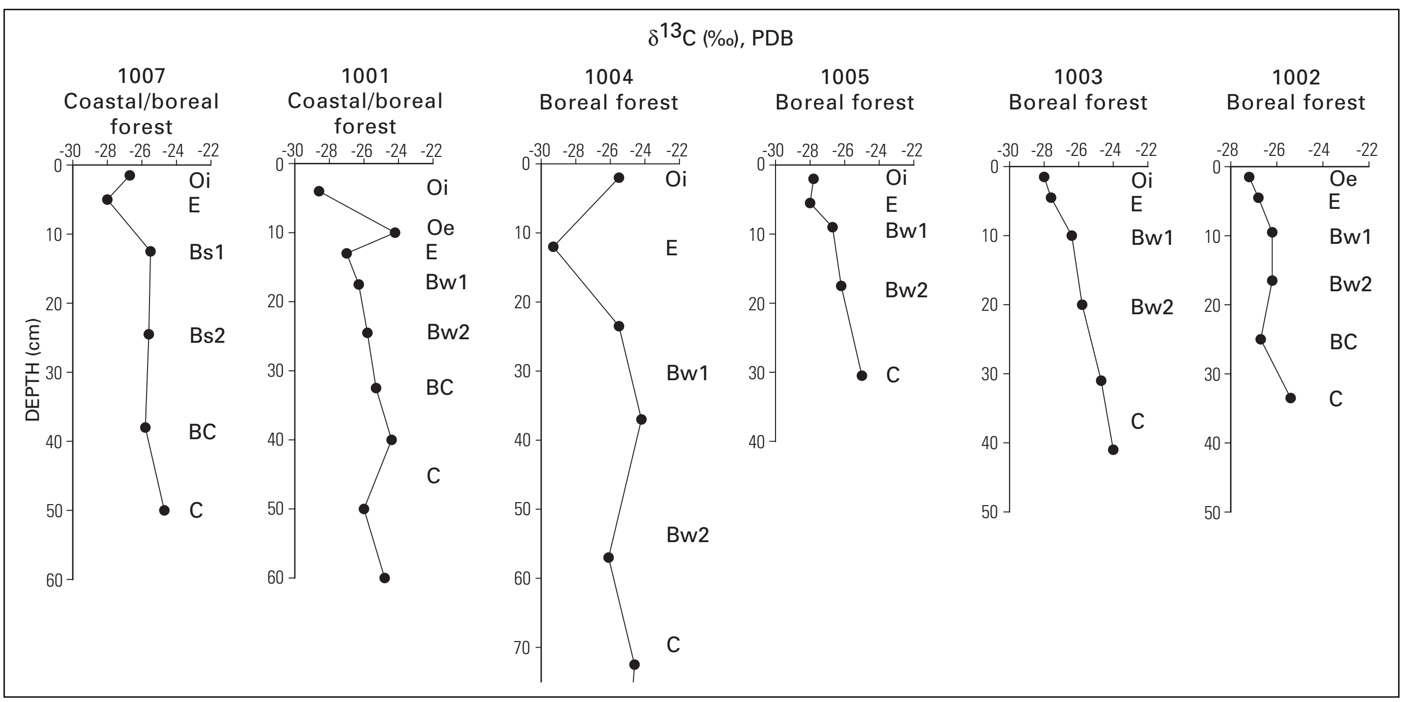

Figure 19. Carbon isotopic composition of soil organic matter vs. depth in Spodosols in the Kenai Peninsula area. Pedon numbers as in figures 6 and 10.

Acknowledgments

Matt Emmons performed the carbon isotopic analyses. We thank James Bockheim, Marith Reheis, and Art Bettis for helpful comments on an earlier draft of the paper.

References Cited

Ager, T.A., 1983, Holocene vegetational history of Alaska, *in* Wright, H.E., Jr., and Porter, S.C., eds., Late Quaternary Environments of the United States, Volume 2: The Holocene: Minneapolis, University of Minnesota Press, p. 128–141.

Ager, T.A., and Brubaker, L., 1985, Quaternary palynology and vegetational history of Alaska, *in* Bryant, V.M., Jr., and Holloway, R.G., eds., Pollen Records of Late-Quaternary North American Sediments: Dallas, Tex., American Association of Stratigraphic Palynologists Foundation, p. 353–383.

Ager, T.A., and Sims, J.D., 1982, Late Quaternary pollen record from Hidden Lake, Kenai Peninsula, Alaska: Palynology, v. 6, p. 271–272.

Anderson, P.M., and Brubaker, L.B., 1994, Vegetation history of north-central Alaska: A mapped summary of late-Quaternary pollen data: Quaternary Science Reviews, v. 13, p. 71–92.

Anderson, P.M., Lozhkin, A.V., Eisner, W.R., Kozhevnikova, M.V., Hopkins, D.M., Brubaker, L.B., and Colinvaux, P.A., 1994, Two late Quaternary pollen records from south-central Alaska: Géographie Physique et Quaternaire, v. 48, p. 131–143.

Ball, D.F., 1964, Loss-on-ignition as an estimate of organic matter and organic carbon in non-calcareous soils: Journal of Soil Science, v. 15, p. 84–92.

Begét, J.E., 1990, Middle Wisconsin climate fluctuations recorded in central Alaskan loess: Géographie Physique et Quaternaire, v. 44, p. 3–13.

Begét, J.E., 1996, Tephrochronology and paleoclimatology of the last interglacial-glacial cycle recorded in Alaskan loess deposits: Quaternary International, v. 34–36, p. 121–126.

Begét, J.E., and Hawkins, D.B., 1989, Influence of orbital parameters on Pleistocene loess deposition in central Alaska: Nature, v. 337, p. 151–153.

Begét, J.E., Stone, D.B., and Hawkins, D.B., 1990, Paleoclimatic forcing of magnetic susceptibility variations in Alaskan loess during the Quaternary: Geology, v. 18, p. 40–43.

Birkeland, P.W., 1999, Soils and geomorphology: New York, Oxford University Press, 430 p.

Bockheim, J.G., and Tarnocai, C., 1998, Recognition of cryoturbation for classifying permafrost-affected soils: Geoderma, v. 81, p. 281–293.

Bockheim, J.G., Tarnocai, C., Kimble, K.M., and Smith, C.A.S., 1997, The concept of gelic materials in the new Gelisol order for permafrost-affected soils: Soil Science, v. 162, p. 927–939.

Bockheim, J.G., Walker, D.A., Everett, K.R., Nelson, F.E., and Shiklomanov, N.I., 1998, Soils and cryoturbation in moist nonacidic and acidic tundra in the Kuparuk River Basin, Arctic Alaska, U.S.A.: Arctic and Alpine Research, v. 30, p. 166–174.

Bozarth, S., 1993, Biosilicate assemblages of boreal forests and aspen parklands, *in* Pearsall, D.M., and Piperno, D.R., eds., Current Research in Phytolith Analysis: Applications in Archaeology and Paleoecology: Philadelphia, The University Museum of Archaeology and Anthropology, p. 95–105.

Brown, J., and Kreig, R.A., 1983, Elliott and Dalton Highways, Fox to Prudhoe Bay, Alaska: Alaska Geological and Geophysical Surveys Guidebook 4, 230 p.

Burt, R., and Alexander, E.B., 1996, Soil development on moraines of Mendenhall Glacier, southeast Alaska. 2. Chemical transformations and soil micromorphology: Geoderma, v. 72, p. 19–36.

Cerling, T.E., Quade, J., Wang, Y., and Bowman, J.R., 1989, Carbon isotopes in soils and palaeosols as ecology and palaeoecology indicators: Nature, v. 341, p. 138–139.

Chapin, F.S., III, Miller, P.C., Billings, W.D., and Coyne, P.I., 1980, Carbon and nutrient budgets and their control in coastal tundra, *in* Brown, J., Miller, P.C., Tieszen, L.L., and Bunnell, F.L., eds., An Arctic Ecosystem: The Coastal Tundra at Barrow, Alaska: Stroudsburg, Pa.: Dowden, Hutchinson & Ross, Inc., p. 458–482.

Coulter, H.W., Hopkins, D.M., Karlstrom, T.N.V., Péwé, T.L., Wahrhaftig, C., and Williams, J.R., 1965, Map showing extent of glaciations in Alaska: U.S. Geological Survey Miscellaneous Geologic Investigations Map I-415, scale 1:2,500,000.

Davies, B.E., 1974, Loss-on-ignition as an estimate of soil organic matter: Soil Science Society of America Proceedings, v. 38, p. 150–151.

Dean, W.E., 1974, Determination of carbonate and organic matter in calcareous sediments and sedimentary rocks by loss on ignition: comparison with other methods: Journal of Sedimentary Petrology, v. 44, p. 242–248.

Everett, K.R., and Brown, J., 1982, Some recent trends in the physical and chemical characterization and mapping of tundra soils, Arctic slope of Alaska: Soil Science, v. 133, p. 264–280.

Everett, K.R., and Parkinson, R.J., 1977, Soil and landform associations, Prudhoe Bay area, Alaska: Arctic and Alpine Research, v. 9, p. 1–19.

Fredlund, G.G., and Tieszen, L.L., 1997, Phytolith and carbon isotope evidence for late Quaternary vegetation and climate change in the southern Black Hills, South Dakota: Quaternary Research, v. 47, p. 206–217.

Gersper, P.L., Alexander, V., Barkley, S.A., Barsdate, R.J., and Flint, P.S., 1980, The soils and their nutrients, *in* Brown, J., Miller, P.C., Tieszen, L.L., and Bunnell, F.L., eds. An Arctic Ecosystem: The Coastal Tundra at Barrow, Alaska, Stroudsburg, Pa.: Dowden, Hutchinson & Ross, Inc., p. 219–254.

Hamilton, T.D., 1978, Surficial geologic map of the Philip Smith Mountains quadrangle, Alaska: U.S. Geological Survey Miscellaneous Field Studies Map MF-879A, scale 1:250,000.

Hamilton, T.D., 1994, Late Cenozoic glaciation of Alaska, *in* Plafker, G., and Berg, H.C., eds., The Geology of Alaska: Boulder, Colorado, Geological Society of America, The Geology of North America, v. G-1, p. 813–844.

Hamilton, T.D., Ager, T.A., and Robinson, S.W., 1983, Late Holocene ice wedges near Fairbanks, Alaska, U.S.A.: Environmental setting and history of growth: Arctic and Alpine Research, v. 15, p. 157–168.

Hamilton, T.D., and Brigham-Grette, J., 1991, The last interglaciation in Alaska: Stratigraphy and paleoecology of potential sites: Quaternary International, v. 10-12, p. 49–71.

Hamilton, T.D., Craig, J.L., and Sellmann, P.V., 1988, The Fox permafrost tunnel: A late Quaternary geologic record in central Alaska: Geological Society of America Bulletin, v. 100, p. 948–969.

Holliday, V.T., 1995, Stratigraphy and paleoenvironments of late Quaternary valley fills on the Southern High Plains: Geological Society of America Memoir 186, 136 p.

Hovan, S.A., Rea, D.K., Pisias, N.G., and Shackleton, N.J., 1989, A direct link between the China loess and marine $\delta^{18}O$ records: Aeolian flux to the north Pacific: Nature, v. 340, p. 296–298.

Janitzky, P., 1986, Organic carbon (Walkley-Black method), *in* Singer, M.J., and Janitzky, P., eds., Field and Laboratory Procedures Used in a Soil Chronosequence Study: U.S. Geological Survey Bulletin 1648, p. 34–36.

Jenny, H., 1980, The soil resource: Origin and behavior: New York, Springer-Verlag, 377 p.

Joint Federal-State Land Use Planning Commission for Alaska, 1973, Major ecosystems of Alaska, scale, 1:2,500,000.

Letkeman, L.P., Tiessen, H., and Campbell, C.A., 1996, Phosphorus transformations and redistribution during pedogenesis of western Canadian soils: Geoderma, v. 71, p. 201–218.

McDowell, P.F., 1997, Stratigraphy and genetic interpretation of a sequence of loess-related deposits in the Yukon Flats southern marginal upland near Circle, Alaska, *in* Edwards, M.E., Sher, A.V., and Guthrie, R.D., eds., Terrestrial Paleoenvironmental Studies in Beringia: Fairbanks, The Alaska Quaternary Center, University of Alaska, Fairbanks, p. 49–51.

Michaelson, G.J., Ping, C.L., and Kimble, J.M., 1996, Carbon storage and distribution in tundra soils of Arctic Alaska, U.S.A.: Arctic and Alpine Research, v. 28, p. 414–424.

Muhs, D.R., Ager, T.A., and Begét, J.E., in press, Vegetation and paleoclimate of the last interglacial period, central Alaska: Quaternary Science Reviews.

Muhs, D.R., Ager, T., Stafford, T.W., Jr., Pavich, M., Begét, J.E, and McGeehin, J.P., 1997, The last interglacial-glacial cycle in late Quaternary loess, central interior Alaska [abs.], *in* Elias, S.A. and Brigham-Grette, J., eds., Program and Abstracts: Beringian Paleoenvironments Workshop, Florissant, Colo., p. 109–112.

Muhs, D.R., Stafford, T.W., Jr., Been, J., Mahan, S., Burdett, J., Skipp, G., and Rowland, Z.M., 1997, Holocene eolian activity in the Minot dune field, North Dakota: Canadian Journal of Earth Sciences, v. 34, p. 1442–1459.

Nelson, D.W., and Sommers, L.E., 1982, Total carbon, organic carbon, and organic matter, *in* Page, A.L., ed., Methods of Soil Analysis: Part 2 Chemical and Microbiological Properties (2nd ed.): Madison, Wis., American Society of Agronomy, p. 539–579.

O'Leary, M.H., 1988, Carbon isotopes in photosynthesis: BioScience, v. 38, p. 328-336.

Orem, W.H., Colman, S.M., and Lerch, H.E., 1997, Lignin phenols in sediments of Lake Baikal, Siberia: Application to paleoenvironmental studies: Organic Geochemistry, v. 27, p. 153–172.

Orem, W.H., Lerch, H.E., and Kotra, R.K., 1993, Lignin oxidation products in sediments of Lake Baikal: Indicators of late Quaternary paleovegetation and paleoclimate change in north-central Asia: Russian Geology and Geophysics, v. 34, p. 89–100.

Péwé, T.L., 1975, Quaternary geology of Alaska: U.S. Geological Survey Professional Paper 835, 145 p.

Péwé, T.L., Berger, G.W., Westgate, J.A., Brown, P.M., and Leavitt, S.W., 1997, Eva Interglaciation forest bed, unglaciated east-central Alaska: Global warming 125,000 years ago: Geological Society of America Special Paper 319, 54 p.

Ping, C.L., Michaelson, G.J., Loya, W.M., Chandler, R.J., and Malcolm, R.L., 1997, Characteristics of soil organic matter in Arctic ecosystems of Alaska, *in* Lal, R., Kimble, J.M., Follett, R.F., and Stewart, B.A., eds., Soil Processes and the Carbon Cycle: Boca Raton, CRC Press, p. 157–167.

Reger, R.D., Pinney, D.S., Burke, R.M., and Wiltse, M.A., 1996, Catalog and initial analyses of geologic data related to middle to late Quaternary deposits, Cook Inlet region, Alaska: State of Alaska Division of Geological and Geophysical Surveys Report of Investigations 95-6, 188 p.

Rieger, S., 1983, The genesis and classification of cold soils: New York, Academic Press, 230 p.

Rieger, S., Schoephorster, D.B., and Furbush, C.E., 1979, Exploratory soil survey of Alaska: Washington, U.S. Government Printing Office, Soil Conservation Service, 213 p.

Runge, E.C.A., Walker, T.W., and Howarth, D.T., 1974, A study of late Pleistocene loess deposits, South Canterbury, New Zealand. Part I. Forms and amounts of phosphorus compared with other techniques for identifying paleosols: Quaternary Research, v. 4, p. 76–84.

Shroba, R.R., and Birkeland, P.W., 1983, Trends in late-Quaternary soil development in the Rocky Mountains and Sierra Nevada of the Western United States, *in* Porter, S.C., ed., Late-Quaternary Environments of the United States, v. 1: The Late Pleistocene: Minneapolis, University of Minnesota Press, p. 145–156.

Smeck, N.E., 1973, Phosphorus: An indicator of pedogenetic weathering processes: Soil Science, v. 115, p. 199–206.

Smeck, N.E., 1985, Phosphorus dynamics in soils and landscapes: Geoderma, v. 36, p. 185–199.

Soil Survey Staff, 1998, Keys to Soil Taxonomy (8th ed.): Washington, D.C., U.S. Department of Agriculture, Natural Resources Conservation Service, 325 p.

Tieszen, L.L., 1991, Natural variations in the carbon isotope values of plants: Implications for archaeology, ecology, and paleoecology: Journal of Archaeological Science, v. 18, p. 227–248.

Ugolini, F.C., Teanier, R.E., Rau, G.H., and Hedges, J.I., 1981, Pedological, isotopic, and geochemical investigations of the soils at the boreal forest and alpine tundra transition in northern Alaska: Soil Science, v. 131, p. 359–374.

Van Cleve, K., Chapin, F.S., III, Dyrness, C.T., and Viereck, L.A., 1991, Element cycling in taiga forests: State-factor control: BioScience, v. 41, p. 78–88.

Van Cleve, K., Dyrness, C.T., Marion, G.M., and Erickson, R., 1993, Control of soil development on the Tanana River floodplain, interior Alaska: Canadian Journal of Forest Research, v. 23, p. 941–955.

Viereck, L.A., and Little, E.L., Jr., 1972, Alaska trees and shrubs: U.S. Department of Agriculture Handbook 410, 265 p.

Walker, D.A., Auerbach, N.A., Bockheim, J.G., Chapin, F.S., III, Eugster, W., King, J.Y., McFadden, J.P., Michaelson, G.J., Nelson, F.E., Oechel, W.C., Ping, C.L., Reeburg, W.S., Regli, S., Shiklomanov, N.I., and Vourlitis, G.L., 1998, Energy and trace-gas fluxes across a soil pH boundary in the Arctic: Nature, v. 394, p. 469–472.

Walker, T.W., and Syers, J.K., 1976, The fate of phosphorus during weathering: Geoderma, v. 15, p. 1–19.

Reviewers: James Bockheim, Marith Reheis, and Art Bettis