III
SURFICIAL GEOLOGY AND SOILS
SURFICIAL GEOLOGY OF THE GRASSLAND AREAS OF BRITISH COLUMBIA AND ADJACENT REGIONS

J.M. Ryder

INTRODUCTION

The purpose of this paper is to provide information about surficial materials and landforms that is pertinent to grassland classification. Surficial material is loose, non-consolidated sediment of relatively young geological age that overlies bedrock and constitutes the parent material for soil. It may consist only of weathered bedrock a few centimetres thick on a high mountain slope, or, at the other extreme, of an interlayered sequence of gravels, silts, and tills extending to depths of several hundred metres below a valley floor.

Five aspects of surficial material are important for any consideration of vegetation characteristics. Firstly, the physical properties of the material, such as texture, compaction, permeability, stratigraphy and landform influence vegetation through their effect upon soil or land surface properties such as soil texture, drainage, depth to impermeable or root-restricting substrate, and microclimate. Secondly, the mineralogical composition (lithology) of surficial material influences soil chemistry and pedological processes. Third, as a result of the first two effects, the spatial distribution of surficial materials may be expected to bear some relationship to vegetation patterns, at least within a given climatic region. A fourth characteristic that is of significance to both soil and vegetation development is the age of the material or the age of the land surface that it constitutes. Finally, to supply a somewhat broader perspective, regional differences, both within British Columbia and further afield, should be considered.
In order to describe these five attributes, I wish to discuss the geological history of the grassland areas. This will provide an explanatory account of the characteristics and distribution of surficial materials which you then will be able to apply to particular areas and specific problems. A bibliography of surficial geology of the grassland areas of British Columbia (Appendix I) and a table of surficial material characteristics (Appendix II) are attached to this paper.

GLACIAL AND POSTGLACIAL HISTORY

Within the past million years, there has been repeated glaciation of North America. Each major episode of cooler climate lasted for about 100,000 years and included several glacier advances and recessions. The most recent interval of generally cooler (although fluctuating) conditions is known in North America as the Wisconsin Glacial Stage (cf. Flint 1971). Since most surficial materials and landforms in Canada are of Wisconsin (chiefly late Wisconsin) age, and since much of this time interval falls within the range of radiocarbon dating, the history of this stage is relatively well known.

It is generally supposed that during each glaciation, two large, discrete ice sheets existed over northern North America (Fig. 1). The Cordilleran Ice Sheet covered the western Canadian mountains and major ice lobes extended southward into Washington, Idaho, and Montana (Fig. 2). The Laurentide Ice Sheet, which originated in uplands east and west of Hudson Bay, extended over most of the remaining glaciated area. At various times during glacial conditions, the ice sheets were contiguous, or nearly so, along the western margin of the Great Plains.

Evidence of two Cordilleran glaciations, both of Wisconsin age, is widespread in British Columbia and the adjacent parts of Washington, Idaho, and Montana (Table 1). Drift* of the late-Wisconsin glaciation, which is known as Fraser Glaciation in British Columbia (Armstrong et al. 1965), and postglacial sediments constitute the present land surface over most of the glaciated area.

*A collective term for till, glaciofluvial, glaciolacustrine, and other glaciogenic materials.
The Quaternary glacial record is longer to the east of the Rockies. Many superimposed till sheets, some of which date back to pre-Wisconsin times, are exposed in cutbanks along deeply incised rivers (cf. Stalker 1976). The chronology of Wisconsin glaciations by both Cordilleran and Laurentide ice sheets is roughly equivalent to that of British Columbia (Table 1).

Figure 1. Generalized map of Pleistocene ice sheets in North America. The western part of the Laurentide Ice Sheet is also known as the Keewatin Ice Sheet (After Flint 1971).
Figure 2. Likely extent of ice in British Columbia, Alberta and adjacent United States during Wisconsin glaciations. Small areas such as the Cypress Hills that were completely surrounded by early Wisconsin ice are not delimited. (Compiled from Richmond et al. 1965, Lemke et al. 1965, Stalker 1977).
Table 1. Chronology of the Wisconsin Glacial Stage in southern British Columbia and adjacent areas

<table>
<thead>
<tr>
<th>REGIONS</th>
<th>SOUTHERN BRITISH COLUMBIA (Fulton &amp; Smith 1983)</th>
<th>SOUTHWESTERN ALBERTA (Barker &amp; Harrison 1977)</th>
<th>CORDILLERAN REGION OF E.WASH., IDAHO, &amp; MONTANA (Richmond et al. 1985)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CHRONOLOGY</td>
<td>~ 10,000 postglacial sediments</td>
<td>postglacial sediments</td>
<td>~ 8,000 postglacial sediments</td>
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<tr>
<td>HOLOCENE</td>
<td>FRASER GLACIATION</td>
<td>WATERTON IV GLACIATION</td>
<td>PINEDALE GLACIATION</td>
</tr>
<tr>
<td></td>
<td>Kemoops Lake Drift</td>
<td>MEDICINE HAT ADVANCE</td>
<td>(3 glacial phases)</td>
</tr>
<tr>
<td></td>
<td>~ 20,000 yrs B.P.</td>
<td>LETHBRIDGE ADVANCE</td>
<td></td>
</tr>
<tr>
<td>QUATERNARY PERIOD</td>
<td>OLYMPIA INTERGLACIATION</td>
<td>interglaciation</td>
<td></td>
</tr>
<tr>
<td>Pleistocene</td>
<td>Beasette Sediments</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Wisconsin Glacial Stage</td>
<td>~ 45,000 yrs B.P.</td>
<td></td>
<td></td>
</tr>
<tr>
<td>(early)</td>
<td>OKANAGAN CENTRE GLACIATION</td>
<td>WATERTON III GLACIATION</td>
<td>BULL LAKE GLACIATION</td>
</tr>
<tr>
<td></td>
<td>Okanagan Centre Drift</td>
<td>ERRATICS TRAIN ADVANCE</td>
<td>(3 glacial phases)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>WATERTON II GLACIATION</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>BUFFALO LAKE TILL</td>
<td></td>
</tr>
</tbody>
</table>

Fraser Glaciation in the Southern Interior of British Columbia

During the onset of Fraser Glaciation, ice accumulation in the mountains proceeded slowly for as long as 10,000 years (Clague 1977). Advancing valley glaciers finally coalesced and extended across the intermontane plateaux shortly after 19,000 to 17,000 years B.P. (Fig. 3a). Subsequent glacier expansion must have been rapid, since the ice sheet reached its maximum thickness and extent slightly after 15,000 years ago. Glacier recession was also rapid; retreat commenced before 13,000 years B.P., and the interior plateaux and valleys of southern British Columbia became ice free during the next two or three millenia. (A summary of radiocarbon-dated glacial history is given in Clague 1981).
Figure 3.  (a) Formation of the Cordilleran Ice Sheet in southern British Columbia.
(b) Deglaciation of southern British Columbia.

Ice Thickness and Flow Directions

At Fraser Glaciation maximum, an unbroken ice sheet, which was possibly an ice dome (cf. Fulton 1975), extended over the Fraser and Thompson plateaux (Fig. 3). Estimates of ice surface elevations (cf. Fulton 1967) indicate that even the higher parts of the plateau lay beneath 500 to 1000 metres of ice. The Coast and Columbia mountains, and the western part of the Rocky Mountains, were occupied by icefields and broad valley glaciers. Only the highest peaks and ridge crests protruded through the general ice cover.

The interior ice sheet, hemmed in by the Coast and Columbia mountains, escaped southward across the Thompson Plateau and northward across the Fraser Plateau (Prest et al. 1967, Fulton 1975, Tipper 1971). Minor diversions from this pattern were controlled by outflow from accumulation zones and underlying topography. For example, ice flowing off the Coast Mountains moved eastward across the Clear Range and upper Hat Creek Valley (Ryder 1976). As the ice became thinner during deglaciation, local topographic controls became more pronounced. For example, ice moved eastward along the Thompson River Valley from Cache Creek junction toward Kamloops at this time, whereas earlier flow had been across the valley toward the south (Fulton 1975). Large trunk glaciers flowed southward along major valleys such as the Rocky Mountain Trench, the Purcell Trench (between the Purcell and Selkirk mountains) and the Okanagan Valley.

Deglaciation

The southern part of the Cordilleran Ice Sheet melted partially by "downwasting and stagnation" and partially by "normal retreat" of glacier termini. It is important to distinguish these two modes of deglaciation, (although they are by no means mutually exclusive) since a distinctive suite of surficial deposits and landforms results in each case.
Downwasting and stagnation predominated in areas of moderate relief, such as the Nicola and Thompson River basins of the southern Thompson Plateau (Fulton 1967, 1975). During relatively rapid climatic amelioration at the end of Fraser Glaciation, snowfall became insufficient to maintain the ice dome over the Interior Plateau; reduced snow accumulation in adjacent mountains also diminished the volume of contributing valley glaciers. Consequently, an undernourished mass of ice remained stranded over the plateau and valleys (Fig. 3b). Emergence of the land surface from beneath this ice can be divided into four phases (Fulton 1967, 1975).

(1) Active ice phase: whilst the thickness of the ice sheet remained greater than the local relief of the land surface, regional ice flow continued with flow direction controlled by the slope of the ice surface. Compact basal till (lodgement till) in a smooth mantle and ice-moulded, streamlined forms such as drumlins, are representative of this phase.

(2) Transitional upland phase: lowering of the ice surface resulted in emergence of the highest uplands. Minor regional flow continued within thick ice where valley alignment coincided with the general ice surface slope. Most meltwater drained over the ice surface, although high-level meltwater channels were cut across emerging ridgecrests that disrupted the supraglacial drainage pattern.

(3) Stagnant ice phase: with the appearance of more upland areas, the ice sheet became subdivided into many isolated masses in valleys and depressions, and regional ice movement ceased. Meltwater flow at this stage was predominantly along the ice margins, giving rise to lateral meltwater channels in parallel arrays, and associated glaciofluvial deposits. Till was either eroded by meltwater or buried beneath sands and gravels.
(4) Dead ice phase: All ice flow ceased when either surface gradients were sufficiently reduced due to local flow or downwasting or the ice became too thin for plastic flow. Meltwater channels at this stage were predominantly subglacial, although some lateral drainage continued and local ponding occurred. The lowering ice surface probably became buried by a considerable thickness of melted-out debris and relatively non-compact meltout till may also have accumulated beneath the ice. Much glacial material was immediately transported and redeposited by meltwater. Thus ridged and hummocky glaciofluvial gravels (eskers, kames), and meltout tills, and pockets of glaciolacustrine sediments typify this final phase of deglaciation.

It should be recognized that neither a general chronology nor a specific relationship between absolute elevation and landform is implied by this mode of deglaciation. For example, the final phase of downwasting may have been underway in a shallow depression on the plateau surface whilst phase 2 was occurring within an adjacent, deep valley. A roughly concentric arrangement of features of phases 3 and 4 may now be found within many topographic basins regardless of their absolute elevation.

**Normal ice retreat** took place where active flow extended to the ice margin. End moraines of various types are indicative of this mode of deglaciation, but these are extremely uncommon in British Columbia. However, small morainal ridges, that may be annual push moraines formed during slight winter readvance of an ice margin, are present on some parts of the Thompson and Fraser plateaux. Examples are to be found in the upper Hat Creek Valley (Aylesworth 1975), near Chapperon Lake east of Douglas Lake (Fulton 1975) and on the plateau surface near the Fraser River in the vicinity of the Gang Ranch.

Other landforms that were commonly exposed by normal ice front retreat, although they were actually formed beneath actively flowing ice, include drumlins, fluted till plains, and mantles of compact basal till (lodgement till). Proglacial rather than lateral meltwater channels, and outwash plains also resulted from this process. Drumlinooid topography is widespread on some parts of the Interior Plateau, for example on the
Fraser Plateau near Riske Creek (Tipper 1971) and on the Thompson Plateau near Separation Lake, south of Kamloops. It is also present within major valleys that were aligned with ice flow direction, for example, the western side of the Thompson River Valley near Ashcroft (Ryder 1976: Fig. 6). The floor of the southern part of the Rocky Mountain Trench is a drumlinized till plain crossed by numerous proglacial meltwater channels (Clague 1975).

Lakes were ponded in many, possibly most, valleys of southern interior British Columbia near the end of Fraser Glaciation. They formed in several distinctive situations:

1) Trunk valleys that drained toward high mountains were blocked by glaciers that still extended from those mountains at a late stage of deglaciation. This gave rise to some of the largest glacial lakes of the region: lakes were impounded in Thompson and Nicola basins by ice flowing eastward from the Coast Mountains (Mathews 1944, Fulton 1969).

2) Other trunk valleys such as the Okanagan were blocked by masses of stagnant or dead ice (cf. Nasmith 1962).

3) Lakes were commonly impounded in tributary valleys by a glacier occupying the trunk valley. For example, Elk, St. Mary and Wigwam River valleys were dammed by a long-lived glacier in the Rocky Mountain Trench (Clague 1975, Ryder 1981: Fig. 2.2).

4) Some lakes were blocked by drift plugs rather than ice. These persisted into postglacial time whilst their outlets were gradually lowered by river erosion. Some of these appear to have been relatively short lived, like Glacial Lake Invermere which was held in the Rocky Mountain Trench by drift near Canal Flats. Others persist to the present day.

5) Isostatic tilting (depression of the land due to the weight of ice, and subsequent rebound upon melting) may also have been responsible for late glacial lake formation and changing lake levels (Fulton 1969).
Postglacial Landscape Development in the Southern Interior of British Columbia

A variety of geomorphological processes has operated over the past 10,000 to 12,000 years to modify the late glacial landscape. Postglacial surficial materials and land surfaces thus range in age from about 12,000 years to less than one year. Land surfaces and hence soils formed on glacial drift may also be younger than the age of their constituent material if they have undergone some erosion during postglacial time.

Fluvial downcutting has occurred along rivers with relatively steep longitudinal gradients and no local base level control, such as a lake, further downstream. Examples include the Fraser and lower Chilecotin rivers and the Thompson River downstream from Kamloops Lake. This erosion has resulted in the formation of canyons, river terraces and scarp slopes. Some of these features have been carved in bedrock, but most commonly they are cut in Fraser Glaciation drift and older unconsolidated materials (Figs. 4 and 5). Terrace surfaces are underlain by fluvial gravels and sands, but these are only a capping that may be very thin upon materials with quite different characteristics. Fluvial terrace cappings commonly overlie glaciolacustrine silts and till within the grassland areas. Scarp slopes typically truncate a horizontally layered sequence of older sediments. Groundwater seepage often occurs on these scarp slopes along contacts between permeable and less permeable materials, for example, where gravels rest upon till.

The highest fluvial terraces constitute relatively old land surfaces that date from early postglacial time and lower terraces are successively younger. However, gradual accumulation of aeolian sand and silt during the postglacial period may have precluded development of deep and mature soil on even the oldest terraces.
Figure 4. Schematic cross-section of the Thompson River Valley at Ashcroft (after Ryder 1976: Fig. 11). Note that a considerable thickness of older drift (till and glaciolacustrine sediments of Okanagan Centre Glaciation - Table 1) was not eroded during Fraser Glaciation. This drift is exposed in the walls of canyons and gullies along tributaries of Thompson River. Preservation of older drift is common within deep valleys that were sheltered from ice erosion, particularly those with a transverse alignment to ice flow direction.

Thompson Valley was partially filled by outwash and deltaic gravels and glaciolacustrine silt during deglaciation. Fluvial terraces were cut into the silt during postglacial degradation by the Thompson River.
Figure 5. Schematic cross-section of Fraser River Valley near Pavilion (after Ryder 1976). Note that Fraser Glaciation till constitutes the highest terrace-like feature in this valley. Lower fluvial terraces have been cut into older drift. Recent landslides in volcanic rocks are widespread along the rim of the Fraser Plateau.
Fluvial deposition has occurred along low-gradient creeks and rivers, and along steeper channels that receive an abundant supply of sediment (chiefly sand and gravel) from upstream. The former situation gives rise to meandering channels; these are typically subject to overbank flooding, which results in deposition of silt layers that are commonly expressed as cumulic soil horizons. The latter situation gives rise to shifting channels that split around bars or vegetated islands and floodplains that consist chiefly of gravel and gravel with a thin capping of sand and silt on relatively undisturbed, vegetated sites. Rivers such as the lower North Thompson and the Similkameen south of Keremeos are of this type. Floodplains constitute the youngest land surfaces, and bear distinctive types and patterns of soil and vegetation as a result.

Alluvial fans have formed during postglacial time where steep tributary creeks emerge onto floodplains or other level surfaces. They consist either of fluvial sand and gravel or of interlayered fluvial and colluvial (debris flow, mudflow) materials. Older alluvial fans rest upon benches and terraces that are now high above river levels, and many have been dissected by their creeks (Ryder 1971). These fan surfaces, like river terraces, may be mantled by aeolian material. Other fans, particularly those that rest upon modern floodplains, are still undergoing sporadic deposition and surface disturbance due to shifting channels. Vaseaux Creek fan in the Okanagan Valley, and Dutch Creek fan at Fairmont in the Rocky Mountain Trench are examples.

Colluvial processes have affected most slopes on both Fraser Glaciation drift and postglacial materials. They include mass movement due to gravity and slopewash by running water.

Mass movement produces many prominent landforms such as landslides, earthflows, colluvial (debris flow) fans and talus slopes. Many of these features stabilized thousands of years ago, but some, such as the Spences Bridge slide, are of recent occurrence. Others are still moving slowly (for example the earthflows at Drynoch near Spences Bridge [VanDine 1980] and at Pavilion [personal communication, 1982, from M. Bovis]
Department of Geography, University of British Columbia, Vancouver, British Columbia] in the Fraser River Valley) or still undergoing sporadic deposition (for example debris flow fans). Thus, materials deposited as a result of mass movement show considerable variation in age as well as in composition and landform.

The process of soil creep, which can be ascribed to both gravity and slopewash, is effective on most sloping land. It results in the development of a mantle of translocated material that overlies (commonly across a transitional zone) undisturbed drift or postglacial sediments. Downslope movement of this mantle may be sufficiently slow that soil development is unaffected, or it may be so rapid that both soil and vegetation are maintained in a perpetually immature state. Terracettes, which are so obvious on most grassland hillsides, are generally considered to be the visible expression of soil creep. The process tends to reduce topographic irregularities by infilling concavities.

Surface erosion by running water has also modified hillslopes in postglacial time, although its effects are not always obvious. Results range from minor sheet erosion to major gully development such as has occurred along the glaciolacustrine silt benches of the South Thompson and Okanagan valleys. It should be noted that where erosion has taken place, the age of the actual land surface may be considerably younger than the age of the underlying material. Strangely enough, surface erosion has been greatest where it is least obvious: many rocky (bedrock and colluvium) hillsides have been completely stripped of their former drift cover. Evidence for this is provided by the apron of fans, composed of reworked drift, that now borders the foot of the slope.

Aeolian processes (wind erosion and deposition) have probably been active throughout the dry interior valleys for most of postglacial time. They were likely most effective during the short interval between deglaciation and the establishment of widespread vegetation. Increased wind activity may have occurred in some areas during a postglacial xerothermic interval (cf. Mathewes and Heusser 1981) if climatic warming resulted in decreased vegetation cover as suggested by Alley (1976) for the Okanagan Valley. Obvious aeolian activity takes place under present day conditions at windy sites. The modern sand dunes at Walhaehin and at Farwell Canyon on the Chilecotin River are well known.
Silt and fine sand particles have been eroded by wind from unvegetated floodplains, lake floors, and steep bluffs (chiefly river banks) of unconsolidated material and deposited on adjacent surfaces. As mentioned earlier, terraces and fans are commonly covered by a distinct veneer of aeolian material. This is generally thickest along the riverward edge of terraces where cliff-top dunes may be present, and it thins toward the valley sides. Aeolian sediment also forms a thin, less distinct capping on till and other drift.

Volcanic eruptions have resulted in deposition of tephra (volcanic ash) across southern British Columbia on several occasions during postglacial time (cf. Westgate et al. 1970). Some ash has been preserved as a conspicuous white layer within postglacial sediments, particularly in fans, talus and aeolian materials. It forms a useful time-synchronous marker bed and provides information about the age of sediments and landforms. Over most drift covered surfaces however, ash has been disseminated within the soil profile where it may have a significant effect upon soil chemistry (cf. Sneddon 1973).

Palynological studies and investigations of glacier behaviour have established that climatic fluctuations occurred within British Columbia and adjacent areas during postglacial time (cf. Mathewes and Heusser 1981, Mack et al. 1979, and Hebda this publication). However, no specific study has been made of the effect of climate change upon landform development or terrestrial sedimentation within southern interior British Columbia. It is possible that changes in precipitation characteristics (intensity and frequency of storms) and in moisture and temperature regimes affected the intensity of surface erosion and resultant deposition of sediment on fans and floodplains.

Glacial and Postglacial History of the Prairie Grassland Areas

The glacial history of southwestern Alberta is complicated by the former juxtaposition of Cordilleran and Laurentide Ice Sheets (Figs. 1 and 2). The temporal
and spatial interrelationships of these two ice masses have been the subject of much
discussion, particularly with regard to the possibility of an ice-free corridor for migration
of early man (summaries in Reeves 1973 and Rutter 1980). During many glaciations,
growth of the two ice sheets appears to have been out of phase: Cordilleran ice reached
its maximum extent and commenced recession before Laurentide ice expanded into the
same area (Stalker and Harrison 1977). Recent work also suggests that each of the
Wisconsin glacial episodes was less extensive than its forerunner. The youngest pair of
overlapping tills are of early Wisconsin age (i.e., older than about 50,000 years). During
late Wisconsin glaciation however, a substantial area of western and southern Alberta and
adjacent Saskatchewan may have remained ice free (Fig. 2) (Stalker 1977, Christiansen
1979, Teller et al. 1980). Thus whilst southern British Columbia lay deep beneath the
Fraser Glaciation ice sheet, parts of the Canadian prairie were ice free and probably
supported tundra vegetation (Mott and Jackson 1982). The surface drift in these areas is
thus several tens of thousands of years older than that elsewhere in the prairies and in
southern British Columbia. (Small areas of relatively old drift also occur in Washington,
Idaho and Montana, Fig. 2).

Many types of drift landforms have been distinguished on the plains. Extensive areas
of hummocky drift - chiefly till but including silts and gravels - formed where parts of the
ice sheet stagnated and melted in place whilst gradually becoming buried by melted-out
debris. Till ridges of a great variety of forms and patterns resulted from processes that
include ice front recession, squeezing of basal till into crevasses, and ice thrusting of
bedrock. Distinctive landforms have been described (for example the "ice-pressed drift
forms" of Stalker [1960]), some of which have not been recognized in British Columbia,
although terminal and recessional moraines are similarly rare. Till plains and fluted
topography are less extensive than the more irregular, hummocky and ridged terrain (see
examples in Gravenor et al. [1960]).
As in British Columbia, large areas were occupied by lakes during deglaciation. Many of these were short lived and left little evidence of their former presence. Others resulted in well defined shoreline features and extensive plains of clay, silt and sand which now constitute the flattest parts of the prairies. Late glacial meltwater cut channels across morainal areas and deposited sand and gravel outwash.

During postglacial time rivers have become entrenched into surficial materials and bedrock. The steep sides of both postglacial and meltwater valleys have been extensively gullied due to surface runoff and wind action. Wind action on outwash and glaciolacustrine plains has commonly resulted in the formation of dunes and deposition of mantles of aeolian sediments (loess). The intensity of postglacial processes, particularly surface runoff and aeolian effects has been influenced by postglacial climate fluctuations (Moran et al. 1976).

THE DISTRIBUTION OF SURFICIAL MATERIALS IN BRITISH COLUMBIA

From the preceding descriptions of geological history, it is apparent that the present day distribution of various types of surficial materials within any given area has resulted from interaction of a number of different factors. The most significant of these are:

(1) Topography, particularly elevation, slope steepness, slope configuration and valley orientation.

(2) Glacial processes, particularly ice thickness, flow direction and mode of deglaciation, including meltwater drainage.

(3) Postglacial processes, particularly river behavior.

Representative transects that show the distribution of surficial materials across several grassland areas are presented in Figures 4 to 8.
Figure 6. Schematic cross-section of the South Thompson River Valley east of Kamloops (after Fulton 1965 and 1967). Outwash and hummocky (kame) gravels were deposited by lateral meltwater streams whilst ice still remained in the valley. Glacial Lake Thompson formed at a later stage of deglaciation and thick glaciolacustrine sediments accumulated. Postglacial downcutting and lateral erosion by the South Thompson River has left prominent silt benches (remnants of the old lake floor) and scarps, and formed low terraces and a floodplain (cf. Fig. 5).
Figure 7. Schematic cross-section of Hat Creek Valley near upper Hat Creek (after Aylesworth 1975). Normal ice-front retreat from the upper slopes of this valley resulted in the formation of small moraine ridges (probably annual push moraines). However, features associated with stagnant and dead ice (hummocky moraine and gravels) occupy the lowest parts of the valley floor.
Figure 8. Schematic cross-section of the Rocky Mountain Trench at the Tobacco Plains (after Clague 1975). Drumlins were formed beneath a large valley glacier that flowed southwards. During deglaciation, they were partly eroded by meltwater and partly buried by outwash sands and gravels.
REFERENCES CITED


Davis, N.F.G. and W.H. Mathews. 1944. Four phases of glaciation with illustrations from southwestern British Columbia. J. Geol. 52:403-413.


APPENDIX I

Bibliography of Surficial Geology for the Grassland Areas of British Columbia

Fraser - CHILCOTIN AREA

Ryder 1976 (see "References Cited")
Tipper 1971 (see "References Cited")


Assessment and Planning Division, B.C. Min. Environ. 1:50,000 scale soils and landform maps for N.T.S. 921 and 92 O. Victoria, B.C.
Thompson – Nicola Area

Aylesworth 1975
Fulton and Smith 1978 see "References Cited"
Mathews 1944
Ryder 1976


Assessment and Planning Division, B.C. Min. Environ. 1:50,000 scale soils and landforms maps for N.T.S. 92 I. Victoria, B.C.

Okanagan – Similkameen Area

Fulton 1969, 1975
Fulton and Smith 1978 see "References Cited"
Nasmith 1962


Assessment and Planning Division, British Columbia Ministry of Environment. 1:50,000 scale soils and landforms maps for N.T.S. 92 L and 1:20,000 soils maps for parts of Okanagan-Similkameen area. Victoria, B.C.

Kettle Valley


East and West Kootenay Areas

Clague 1975 see "References Cited"

Ryder 1981 see "References Cited"


Assessment and Planning Division, B.C. Min. Environ. 1:50,000 scale soil maps and terrain maps for N.T.S. 82G and 82J. Victoria, B.C.
<table>
<thead>
<tr>
<th>MATERIAL CHARACTERISTICS</th>
<th>FLUVIAL SEDIMENTS</th>
<th>COLLUVIUM</th>
<th>AEOLIAN SEDIMENTS</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mode of deposition</td>
<td>-transported by streams &amp; rivers and deposited in channels or on floodplains</td>
<td>-accumulates on or at the foot of slopes as a result of gravity-induced movement of surficial material or bedrock fragments</td>
<td>-transported and deposited by wind</td>
</tr>
<tr>
<td>Texture</td>
<td>-gravel, sand &amp; less commonly silt; moderately to well-sorted</td>
<td>-generally poorly sorted, but particle size &amp; shape dependant upon nature of “parent material”</td>
<td>-coarse silt &amp; sand; well-sorted</td>
</tr>
<tr>
<td>Shape of coarse particles</td>
<td>-subrounded to well-rounded</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Structure</td>
<td>-horizontal or dipping (cross-bedding) stratification</td>
<td>-debris &amp; mudflow deposits (e.g. colluvial fans), slopewash &amp; talus materials are stratified parallel to the land surface; other types of colluvium are generally non-stratified</td>
<td>-mantles are usually non-stratified; dune sands may have dipping stratification &amp; ripple marks</td>
</tr>
<tr>
<td>Compaction</td>
<td>none to slight</td>
<td>-generally non-compact, but some exceptions</td>
<td>none</td>
</tr>
<tr>
<td>Cohesion</td>
<td>-gravel &amp; sand - none; silt - slight to moderate</td>
<td>-increases with increasing silt and clay content of the material</td>
<td>-none to slight; unless cemented by carbonate or gypsum accumulation</td>
</tr>
<tr>
<td>Permeability</td>
<td>-gravel &amp; sand - high; silt - moderate</td>
<td>-varies with texture of material</td>
<td>high</td>
</tr>
<tr>
<td>Typical landforms</td>
<td>-fluvial plains, river terraces, alluvial fans</td>
<td>-talus slopes, hummocky &amp; ridged topography (rockslides, earthflows, stumps) colluvial fans; smooth mantle on slopes</td>
<td>-mantles on terrace fans and other surfaces -dunes (low hillocks &amp; ridges)</td>
</tr>
<tr>
<td>Mineralogical composition</td>
<td>-highly variable except in localized areas where sediments derived from bedrock or till</td>
<td>determined by type of bedrock or other “parent material”</td>
<td>-most commonly quartz and feldspar</td>
</tr>
</tbody>
</table>
# APPENDIX II

Characteristics and Properties of Surficial Materials

*Note:* This table is intended only as a general guide to the most typical characteristics of the common surficial deposits of the grassland areas. Readers should consult the references listed or the bibliography (Appendix I) for more specific descriptions of local materials.

**Definitions:**
- **coarse particles**: particles 2 mm diameter, including gravel, pebbles, cobbles, boulders, stones, etc.
- **fines**: sand, silt, clay.

## TILL (Morainal material)

<table>
<thead>
<tr>
<th>MATERIAL CHARACTERISTICS</th>
<th>Flow till</th>
<th>Ablation (Supraglacial) till</th>
<th>Ablation meltout till</th>
<th>Lodgement till</th>
<th>Basal (subglacial) till</th>
<th>Basal meltout till</th>
<th>Deformation till</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mode of deposition</td>
<td>-flowage of saturated, supra-glacial debris &amp; accumulation either in depressions on the glacier or on adjacent land surface</td>
<td>-accumulates in place on top of glacier due to melting, from the surface downward, of debris-rich ice</td>
<td>-accumulates under pressure at the base of an actively flowing glacier</td>
<td>-accumulates as debris-rich ice melts from the base upward due to geothermal heating; most typically from stagnant or &quot;dead&quot; ice</td>
<td>-pre-existing sediments or bedrock that have been transported &amp; deformed but not totally re-worked, by a glacier</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Texture</td>
<td>-coarse particles in a matrix of fines</td>
<td>-coarse particles in a matrix of fines, but less silt &amp; clay than in other types of till</td>
<td>-coarse particles in a matrix of fines</td>
<td>-same as pre-existing sediments</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Shape of coarse particles</td>
<td>-variable; faceted and striated particles are less common than in basal tills &amp; may be absent</td>
<td>-variable; particles may be faceted (one or more flat faces) &amp; striated</td>
<td>-same as pre-existing sediments</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Structure</td>
<td>-stratified; commonly interlayered with glaciofluvial or glaciolacustrine sediments</td>
<td>-non-stratified</td>
<td>-generally non-stratified; faint stratification may be visible in large exposures</td>
<td>-non-stratified</td>
<td>-some preglacial structures (e.g. stratification) will be preserved, but deformed (e.g. faulted, folded) due to glacial effects</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Compaction</td>
<td>moderate</td>
<td>none to slight</td>
<td>high</td>
<td>slight to moderate</td>
<td>variable</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Cohesion</td>
<td>moderate to high</td>
<td>none to moderate</td>
<td>high</td>
<td>moderate</td>
<td>variable</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Permeability</td>
<td>low to moderate</td>
<td>moderate</td>
<td>impermeable to low</td>
<td>low</td>
<td>variable</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Typical landforms</td>
<td>-hilly, detailed, &amp; irregular topography</td>
<td>-hills; drumlins &amp; fluted topography; smooth mantles on hillsides</td>
<td>-hilly, detailed &amp; irregular topography</td>
<td>-variable</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mineralogical composition</td>
<td>-depends primarily upon type of bedrock (or older sediments) from which till was derived; composition (particularly of basal tills) may vary abruptly over short distances in accord with changes in type of bedrock</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

*Note:* Texture of till will also vary in accord with type of bedrock or older sediments from which till was derived.
# APPENDIX II
Continued

<table>
<thead>
<tr>
<th>MATERIAL CHARACTERISTICS</th>
<th>Glaciofluvial Sediments</th>
<th>Proglacial</th>
<th>Glaciolacustrine Sediments</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Mode of deposition</strong></td>
<td>-transported by meltwater &amp; deposited in contact with ice; either on, within, beside or beneath a glacier</td>
<td>-transported by meltwater &amp; deposited in front of (i.e. downstream from) the ice margin</td>
<td>-accumulate in glacial (ice-dammed) lakes, chiefly by settling of suspended sediment</td>
</tr>
<tr>
<td><strong>Texture</strong></td>
<td>-extremely variable (silt to boulders); poorly sorted</td>
<td>-variable (silt to boulders); poor to moderate sorting</td>
<td>-silt, clay &amp; fine sand on lake floors; sand &amp; gravel in deltas</td>
</tr>
<tr>
<td><strong>Shape of coarse particles</strong></td>
<td>-subangular to subrounded</td>
<td>-subangular to subrounded</td>
<td>-subangular to subrounded</td>
</tr>
<tr>
<td><strong>Structure</strong></td>
<td>-may or may not be stratified; typically displays faults, disturbed stratification &amp; other deformation structures</td>
<td>-horizontal or dipping (cross-bedding) stratification is common, but may be non-stratified</td>
<td>-horizontal or gently undulating stratification in silts &amp; clays; sands may show dipping stratification (cross-bedding) or ripple marks; deltas consist chiefly of long, sloping strata (torset beds)</td>
</tr>
<tr>
<td><strong>Compaction</strong></td>
<td>none to slight</td>
<td>none to slight</td>
<td>none to slight</td>
</tr>
<tr>
<td><strong>Cohesion</strong></td>
<td>-sands and gravels - none; silt or silt matrix - slight</td>
<td></td>
<td>-sand &amp; gravel - none; silt &amp; clay - moderate to strong</td>
</tr>
<tr>
<td><strong>Permeability</strong></td>
<td>moderate to high</td>
<td>high</td>
<td>-sand &amp; gravel - high; silt &amp; clay - low to impermeable</td>
</tr>
<tr>
<td><strong>Typical landforms</strong></td>
<td>-hummocky, ridged &amp; irregular topography (kames &amp; eskers); irregular depressions (kame terraces)</td>
<td>-outwash plains, pitted (kettled) outwash plains; meltwater channels</td>
<td>-gullied beaches &amp; scarps; lacustrine plains; delta-terraces</td>
</tr>
<tr>
<td><strong>Mineralogical composition</strong></td>
<td>-probably similar to local till</td>
<td>-probably similar to till in areas where meltwater originated</td>
<td>-fines probably similar to matrix of local till</td>
</tr>
</tbody>
</table>

**Notes:**
- *ice-contact glaciofluvial sediments commonly contain pockets of till and glaciolacustrine material.*
- *under the E.L.U.C.S. 1976 Terrain Classification System, deltas formed in glacial lakes are grouped with glaciofluvial sediments.*
CHERNOZEMS:

THEIR CHARACTERIZATION AND DISTRIBUTION

A. Green and
A.L. van Ryswyk

INTRODUCTION

By area the Chernozems account for a small portion of the world's soils (Fig. 1). Nevertheless, they are extremely important to agriculture because they can be so tremendously productive for a wide range of crops. Consequently these soils have been surveyed, researched and documented in many publications both in Canada and elsewhere. It is our intention in this paper to characterize Chernozemic soils and to show their distribution by discussing their historical background and how they have been defined in the Canadian, and United States Systems of Soil Classification and the FAO/UNESCO Legend for the Soil Map of the World. We will also review certain studies in British Columbia relating climate to Chernozemic soils.

HISTORICAL BACKGROUND AND CHARACTERISTICS OF CHERNOZEMS

The name Chernozem originally came from the Russian words "chern" meaning black and "zemlja" meaning earth — soils rich in organic matter having a black colour (Fitzpatrick 1980). The Russians were able to observe vast expanses of such soils developed under grasses on the steppes. Focusing attention on the soil profile led them to realize that many soil characteristics are profoundly affected by differences in environmental factors and that there is a fairly close relationship between soils and
vegetation, but more particularly between soil and climate. These observations revolutionized soil science and led to the concept of zonal soils. A zonal soil has a morphology that reflects the influence of the climate and living organisms, mainly vegetation, characteristic of a large area or zone (Canada Department of Agriculture 1976). It was natural that this concept of zonal soils would be adapted by other countries such as Canada and the United States that have great expanses of land (Baldwin et al. 1938).

Figure 1. World distribution of Mollisols (mainly Chernozemic soils).
1. The North American Great Plains
2. From the western side of the Black Sea eastward
3. Manchuria
4. The Argentine Pampas

Chernozems are grassland soils which occur in intermediate regions between arid and humid climates. In the United States the name Chernozem was adopted and defined as a zonal group of soils having deep, dark to nearly black surface horizons that are rich in organic matter and grade into light colored soils below. At 45 to 120 cm, these soils have horizons of accumulated calcium carbonate. They develop under tall and mixed grasses in a temperate to cool, subhumid climate.

Chestnut and Prairie soils are also considered to be grassland soils, occurring respectively in climates drier and moister than the Chernozems. The Chestnut soils develop under low-grass steppe in semiarid regions and are differentiated from Chernozems by their dark brown surface horizon, slightly lower organic matter content and a shallower depth to calcium carbonate in the profile. Prairie soils (later called Brunizems in the United States) have a very dark greyish-brown surface horizon, somewhat deeper profile and greater depth to carbonates than Chernozems. These soils develop under tall grass vegetation in a temperate, humid climate. The Prairie soils or Brunizems are considered to be among the most productive soils in the world.

The map in Figure 2 shows the extent and relationship of the Chernozemic and Prairie zonal soil groups in the United States. The north-south line separating these two soil groups is also the line which separates soils that have a carbonate layer (PEDOCALS) from those that did not have one (PEDALFERS). It is interesting to note that this line is very similar to the line that separates the subarid and arid areas to the west from the humid areas to the east in the United States.

In order to accommodate the growing knowledge of soil science the Americans devised a new system of soil classification (Soil Survey Staff 1975). Many old familiar names such as Chestnut and Chernozem were dropped and the soils were placed in the new order of Mollisols. Chestnut soils came under the suborder of Ustolls, Brunizems became Udolls, and Chernozems became Borolls.
Figure 2. General distribution of the important zonal soil groups of the United States. Reprinted by permission from Soils, Their Origin, Constitution and Classification, Third Edition by G.W. Robinson, (c) 1951 by George Allen and Unwin Ltd., Woodbridge Press, Ltd., Great Britain.
In Canada the concept of the Light Chestnut, Chestnut and Chernozem from abroad was accepted, but, due to uncertainty as to the degree of correlation between Canadian soils and those in other countries, the names assigned to them were Brown, Dark Brown, and Black Chernozems (Leahey 1966). The term Chernozem was used at the order level to describe most of the soils in the grassland and parkland areas of the cool, subarid to subhumid Interior Plains of western Canada. These soils have dark coloured mineral-organic surface horizons and brownish, usually prismatic subsurface non-saline horizons lying on calcareous parent material. They are well saturated with bases. The sola are well drained and free of soluble salts. The order includes four great groups, the Brown, Dark Brown, Black and Dark Gray. The division of great groups is based largely on the measurable differences in the colour of the surface soil which is a reflection of the climate and vegetation under which these soils have developed. It is of interest to note that the Dark Gray great groups came into being after the adoption of the Munsell Soil Colour System. The great groups of this order are defined on the basis of Munsell values of the dry chernozemic surface soil and not on the chroma, although usually the Brown and the Dark Brown soils have a higher chroma than the Black and Dark Gray soils (Leahey 1963).

Specifically, in the Canadian System of Soil Classification, all soils of the Chernozemic order have an A horizon at least 10 cm thick in which the organic matter accumulates (Ah, Ahe, Ap). The A horizon must have a chroma less than 3.5 moist and a colour value darker than 5.5 dry and 3.5 moist and at least one Munsell unit darker than that of the IC horizon. When disturbed by cultivation or other means the A must be thick and dark enough to provide 15 cm of surface material that meets this colour criteria. Additional properties of the A horizon are: it contains 1-17% organic C and has a C:N ratio less than 17; it usually has sufficiently good structure so that it is neither massive and hard nor single grained when dry; it has a base saturation (neutral salt) more than 80%
and calcium as the dominant exchangeable cation, and it has a mean annual temperature of $0^\circ$C or higher and a soil moisture subclass drier than humid. Chernozemic soils may have an Ae and a Bm or a Bt horizon (Canada Soil Survey Committee 1978).

Chernozemic soils do not have the following: solonetzie B, podzolic B, evidence of gleying strongly expressed to meet the criteria of gleysolic soils, or permafrost within 3 metres of the surface (Canada Soil Survey Committee 1978).

Although the chernozemic A horizon used in the Canadian System of Soil Classification closely resembles the mollie epipedon now used in the United States (Soil Survey Staff 1975) and the mollie A horizon used in the legend for the soil map of the World (FAO 1974) they are not synonymous. Note that while an epipedon is an horizon formed at the surface it may be thinner than the chernozemic A horizon or include some of the B horizon.

Specifically, a mollie epipedon is defined by the Soil Survey Staff (1975) as a surface horizon that, when mixed to a depth of 17.5 cm (7 inches), contains 1% or more organic matter with colour values darker than 5.5 dry and 3.5 moist. Structure cannot be both massive and hard or very hard when dry. Base saturation is over 50% and the epipedon is not naturally dry more than 3 months of the year.

The definitions and criteria established for soils in the Chernozemic order are such that they are similar to those in the suborder of Borolls in the United States taxonomy and with Kastanozems, Chernozems and Greyzems in the soil units of the FAO/UNESCO World System. The relationships between the Canadian Chernozemic great soil groups and the equivalents in the United States and FAO/UNESCO systems are shown in Table 1.
Table 1.
Comparison of the Canadian Chernozemic Great Soil Groups and their near equivalents

<table>
<thead>
<tr>
<th>Great Soil Group</th>
<th>Equivalent in U.S. Soil Classification</th>
<th>Equivalent in FAO/UNESCO Classification</th>
<th>Climate</th>
<th>Vegetation</th>
<th>Associated Mainly With Other Great Soil Groups</th>
</tr>
</thead>
<tbody>
<tr>
<td>Brown</td>
<td>Aridic Haploboroll</td>
<td>Haplic Kastanozem</td>
<td>Boreal</td>
<td>Short Grass and Mixed Prairie Grasslands; Sagebrush-Bluebunch Wheatgrass of B.C.</td>
<td>Dark Brown Solonetz Solodized Solonets Solod to a lesser extent Humic Gleysols</td>
</tr>
<tr>
<td></td>
<td>Subarid Soil</td>
<td></td>
<td>Soil Climates</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Boreal and Cytoboreal Soil</td>
<td></td>
<td>Boreal and Cytoboreal Soil Climates Semiard</td>
<td>Midgrass Sections of the Mixed Prairie Grasslands; Bluebunch Wheatgrass - Fescue of B.C.</td>
<td>Brown Black Solonetz Solodized Solonetz and Solod to a lesser extent Humic Gleysols</td>
</tr>
<tr>
<td></td>
<td>Subhumid</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>Rendzina</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Black</td>
<td>Udic Haploboroll</td>
<td>Haplic Chernozem</td>
<td>Cold Cryo-Boreal Sub-Humid to Moderately Cool Boreal Subhumid</td>
<td>Mesophytic Grasses and Forbs Characteristic of the Fescue Prairie Parklands and True Prairie</td>
<td>Dark Browns Dark Grays Humic Gleysols Solonetz Solodized Solonetz and Solod</td>
</tr>
<tr>
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<tr>
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<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Dark Gray</td>
<td>Boralic Boroll</td>
<td>Greyzem</td>
<td>Cold to Moderately Cold Cryo-Boreal Subhumid to Humid Regimes</td>
<td>Mixed Native Vegetation of Tree, Shrubs, Forbs and Grasses in Forest-Grassland Transition Zones</td>
<td>Dark Gray Lavsols Blacks Solonetz Solodized Solonetz and Solod</td>
</tr>
</tbody>
</table>


DISTRIBUTION OF CHERNOZEMS IN CANADA

Strong resemblances exist between the boundaries for the Chernozemic great soil groups (Fig. 3) and the various types of grassland vegetation (Table 1). There is a close similarity between the pattern of great soil groups and soil moisture subclasses for the prairie grasslands as well as for soil temperature classes, although the latter is not as evident (Clayton et al. 1977).

Figure 3. The major areas of Brown, Dark Brown and Black Great Soil Groups in Canada (Legget 1965).
Table 2 shows the extent of each of the four Chernozemic great soil groups.


<table>
<thead>
<tr>
<th>GREAT SOIL</th>
<th>PERCENT OF TOTAL LAND AREA IN CHERNOZEMS IN CANADA</th>
<th>DOMINANT COMPONENT MAP UNITS KM² (square mile)</th>
<th>SUBDOMINANT COMPONENT IN MAP UNITS</th>
</tr>
</thead>
<tbody>
<tr>
<td>BROWN</td>
<td>1.1</td>
<td>100,230 (38,714)</td>
<td>1,760 (680)</td>
</tr>
<tr>
<td>DARK BROWN</td>
<td>1.2</td>
<td>111,044 (42,891)</td>
<td>3,042 (1,175)</td>
</tr>
<tr>
<td>BLACK</td>
<td>2.2</td>
<td>200,655* (77,503)</td>
<td>2,822 (1,090)</td>
</tr>
<tr>
<td>DARK GRAY</td>
<td>0.6</td>
<td>56,181 (21,700)</td>
<td>7,974 (3,080)</td>
</tr>
<tr>
<td>TOTAL</td>
<td>5.1</td>
<td>468,110 (180,808)</td>
<td>15,598 (6,025)</td>
</tr>
</tbody>
</table>

* In addition about 2,200 km² (850 sq. mi.) of these soils may be found in local areas within the southern interior of the Cordilleran Region of British Columbia.

The main areas of Chernozemic soils occur within the prairie and parkland regions of Manitoba, Saskatchewan and Alberta. Estimates of their extent are given in Table 3.

<table>
<thead>
<tr>
<th>SOIL ORDER</th>
<th>GREAT GROUP</th>
<th>MAN.</th>
<th>SASK.</th>
<th>ALTA.</th>
<th>TOTAL</th>
</tr>
</thead>
<tbody>
<tr>
<td>CHERNOZEMIC</td>
<td>BROWN</td>
<td>5,625</td>
<td>3,240</td>
<td>8,865</td>
<td></td>
</tr>
<tr>
<td></td>
<td>DARK BROWN</td>
<td>6,812</td>
<td>3,640</td>
<td>10,452</td>
<td></td>
</tr>
<tr>
<td></td>
<td>BLACK</td>
<td>5,263</td>
<td>6,539</td>
<td>11,802</td>
<td></td>
</tr>
<tr>
<td></td>
<td>DARK GRAY</td>
<td>2,632</td>
<td>1,595</td>
<td>4,227</td>
<td></td>
</tr>
<tr>
<td></td>
<td>TOTAL</td>
<td>7,895</td>
<td>20,571</td>
<td>13,150</td>
<td>41,616</td>
</tr>
</tbody>
</table>

* Estimates from respective soil survey groups.

In comparison the extent of Chernozemic soils in British Columbia is very small and it is difficult to portray them on small scale maps (Fenger this publication).

SOME ENVIRONMENTAL FACTORS RELATING CLIMATE TO CHERNOZEMIC SOILS IN BRITISH COLUMBIA

The fact that pedologists can observe grassland soils in the field and put boundaries between them on the basis of difference in the colour of the chernozemic Ah horizons is helpful in establishing climatic regimes and subclasses where meteorological data is scarce or non-existent. Soil boundaries also help verify significant changes in the vegetation composition.

In British Columbia the vertical succession of soil and vegetation zones is obvious. It is primarily due to the differences in climate accompanying changes in elevation and latitude. This vertical zonation is clearly shown in the Ashcroft map area (see Fenger this publication; Fig. 1).
The relationships among native vegetation, soil organic matter (carbon), certain climatic factors and elevation were first described by Spilsbury and Tisdale (1944). Later they were studied in greater detail across a series of Brown, Dark Brown and Black Chernozemic soils on the Lac du Bois range, near Kamloops, British Columbia (van Ryswyl et al. 1966). The results from the Lac du Bois range study are summarized below.

The gradient of soil organic carbon increases with elevation, 345 to 975 m (Fig. 4). It is low, but consistent, through the Brown and into Dark Brown soils, where it begins to increase more sharply with elevation. Throughout the Black soils, the gradient continues to increase dramatically. This increase is coincident with increasing numbers of plant species and vegetation production.

![Graph showing organic carbon and P/E ratio against elevation.]

Figure 4. Soil organic carbon (OC) in top 2 dm (8 in) and precipitation-evaporation (P/E) ratio across a series of Chernozemic soils on Lac du Bois range near Kamloops, B.C. Locations of weather stations are indicated by solid vertical lines.
Climatic measurements for five months (April to October) averaged over three years, showed that mean air temperature ranged from 15.5°C to 11.5°C and precipitation from 136 to 188 cm across the series of soils. Both climatic factors were generally related to soil organic carbon. However, precipitation to evaporation (P/E) ratio, calculated monthly by Thornthwaite's (1931) method, and averaged over the same three years, integrates these two factors. P/E ratio parallels soil organic carbon (Fig. 4) more closely than does precipitation alone. Climatic moisture surplus/deficit maps have since been prepared (B.C. Ministry of Environment 1980) which indicate the moisture deficit ranges from -400 mm of water at 450 m elevation to -200 mm at 950 m.

There is a sharp break, along this series of Chernozemic soils, between grassland and Douglas-fir (Pseudotsuga menziesii) forest vegetation at about 950 elevation, the position where soil organic carbon and P/E gradients are the greatest. The grassland-forest boundary appears to be quite stable with few tree seedlings successfully establishing beyond the forest edge. It appears this may be the position below which the upper soil layers become too dry for some period in the summer to allow successful tree establishment under contemporary climate.

The soils under the adjacent forest have been classified as Orthic Gray Luvisols (Green and Leskiw 1971). Significant inclusions of Orthic Dark Gray Luvisolic and Orthic Dark Gray Chernozemic soils occur within areas mapped as Orthic Gray Luvisols, but not in areas mapped as Orthic Black soils. This would indicate that grassland had occupied sites (Dark Gray) within the present adjacent forest at some time in the past.

Jakoy (1981) studied soils along three transects that crossed the grassland-forest transition boundary at elevations between 850 to 930 m, also on the Lac du Bois range. He found the grassland soils are typical of the Dark Brown soils, although the ones at
higher elevations have characteristics approaching those of Black soils. The forested soils were classified as Orthic or Eluviated subgroups of Eutric and Melanic Brunisols, although one pedon at the highest elevation approached a Luvisol. The grassland-forest boundary coincided with a change in superficial geologic materials. The grasslands are on compact morainal material or on coarse fluvial material over compact morainal material, whereas the adjacent forests are established on colluvial and glaciofluvial materials. Jackoy feels that tree establishment in these park-like ecotones depends on the presence of materials that are loose enough not to restrict penetration of tree roots, especially ponderosa pine (Pinus ponderosa), and that have better moisture regimes than those of adjacent materials.

Very few studies have been done on the moisture regimes of British Columbia grassland soils. However, soil water content (gravimetric) and soil water potential (thermocouple psychrometer) were measured over the growing season at a Black and a Dark Brown soil site both within 20 km of the Lac du Bois range (unpublished data, Kamloops Research Station, Agriculture Canada). Both sites were occupied by a stand of introduced crested wheatgrass (Agropyron desertorum) in good to excellent condition. The Black soil site was on compact till overlain by 50 cm of loamy aeolian and morainal material. It was adjacent to Douglas-fir forest at 1035 m elevation. The Dark Brown soil site at 610 m was at the toe of a series of coalescing fans on 50 cm of loam over sandy loam to loamy sand. About 300 m lateral distance upslope on the coarser textured apex of the fan, soil mapped as Degraded Eutric Brunisol (Young 1978) supported a stand of open ponderosa pine and bluebunch wheatgrass (Agropyron spicatum).

Soil water potential on the Black soil site, reached -15 bar at the 50 cm depth by mid to late July. This coincided with cessation of growth of crested wheatgrass even though soil water potential at 100 cm in the compact till did not fall below -10 bar at any time during the growing season. Few fine grass roots penetrated below 50 cm.
At the Dark Brown site, crested wheatgrass growth also ceased when water potential reached -15 bar, but this occurred about one month earlier than it did at the Black soil site. Water potential at 100 cm was less than -15 bar throughout the entire year.

No soil moisture measurements were made under the ponderosa pine stand adjacent to the Dark Brown site. However, one might speculate that the coarser materials under the ponderosa pine have a lower water holding capacity than do those of the Dark Brown site and would, therefore, allow precipitation to penetrate deeper into the profile. Water in the lower parts of the profile is less subject to loss by evaporation from the soil surface than is water nearer the surface. The lower water is available to the deep rooted ponderosa pine, allowing the pine to survive summer drought.

The transition between forest and grassland appears to be sensitive to plant available moisture. Factors that affect moisture availability, such as water holding characteristics of surficial materials, slope and direction of exposure and elevation, become very significant in determining vegetation type.

These factors have been recognized by Young (1978) in mapping the soils of the Ashcroft area. He indicates that Black and Dark Brown Chernozemic soils are associated with Douglas-fir and/or ponderosa pine, but that grassland vegetation is determined by variations in soil materials (edaphic) as well as by past fire and grazing history. Brown soils, however, were associated only with grassland vegetation of the interior bunchgrass zone (Young 1978).

McLean and Green (1979) have shown the relationship among elevation, soil great group, soil parent material and habitat type within plant zones for the Princeton area. Ponderosa pine-Idaho fescue (Festuca idahoensis) habitat type occurs on Dark Gray
Chernozem soils on coarse outwash, colluvium and ponded deposits on terraces and fans. Big basin sagebrush (*Artemisia tridentata*), bluebunch wheatgrass and big basin sagebrush-Idaho fescue habitat types are on Dark Brown and Dark Gray Chernozemic soils respectively, but both are on glacial till. Idaho fescue-parsnip-flowered umbrella-plant (*Eriogonum heracleoides*) habitat type is found on Black, Dark Brown or Dark Gray soils on glacial till or colluvium. These same materials associate with Rocky Mountain Douglas-fir-bluebunch wheatgrass habitat type on Dark Gray soils.


Relationships have been observed in the field between various subgroups of Chernozemic, Brunisolic or Luvisolic soils and grassland or forest vegetation. For example, forage production of vegetation in good to excellent condition increases from Brown through Dark Brown to Black Chernozemic soils (McLean and Marchand 1968). There are strong indications that surficial geological materials and topography greatly influence the distribution of grassland and forest vegetation, particularly at the grassland-forest ecotone, through their influence on soil water holding characteristics. More field measurements of seasonal soil water regime within the root zone, at strategically located sites, are needed to confirm these observations. They should be designed to show how differing water regimes favour one vegetation type over another and the extent to which they exert influence in different climatic regions.
In summary it can be said that the area of Chernozemic soils is small compared to the total area of Canada, however, they are extremely valuable to agriculture. Chernozemic soils comprise about 80% of the cropland and a large proportion of the rangeland. They produce a great variety of crops ranging from cereals and tree fruits to forages and livestock.

These soils have been surveyed, studied, described, tested and documented in numerous reports, maps and research papers both in Canada and elsewhere. New avenues for exploration perhaps lie in the grassland-forest transition where the Chernozemic soils may meet the Luvisols, Brunisols or Podzols. In British Columbia there is a need for field measurement of the soil water regime to confirm relationships among grassland and forest vegetation, soil great groups and unconsolidated geological materials. Competition between wildlife and domestic livestock for alpine range will undoubtedly increase, and more information about Black soils and how they merge with the soils of the subalpine parkland is needed if the fragile ecology of these areas is to be managed successfully.

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SOILS OF GRASSLANDS AT CLIMATIC CLIMAX:
THEIR PROCESSES AND DYNAMICS

S. Pawluk

INTRODUCTION

Soils of the grassland region are characterized by ubiquitous features that reflect a degree of uniqueness within their environment. Namely, the development of a chernozemic A horizon dominates all landscapes.

The chernozemic A horizon is an organically enriched surface soil layer that varies in colour from brown through black to dark gray depending upon the kind and amount of organic matter present. It is usually relatively thick, friable and possesses well developed granular structure. Chemically, the horizon is generally nutrient-rich with more than 80 percent of the exchange surface occupied by calcium ions (Canada Soil Survey Committee 1978). The humus of Chernozemic soils originates from the steppe grasses with which it is associated.

Salient features that define the chernozemic A horizon provide growing conditions that make these soils among the most naturally fertile in the world. Physical characteristics reflecting good structure and tilth coupled with high native nutrient status provide a rooting environment highly attractive to a wide variety of annual and perennial crops. Thus they produce about 90 percent of the grain in commercial trade in addition to forage for the livestock industry (United States Department of Agriculture 1957). The capability of grassland soils to sustain their productive capacity depends a great deal on the continuance of the processes involved in their formation.
FACTORS OF SOIL FORMATION IN THE GRASSLANDS

Processes involved in the formation of Chernozemic soils of the grassland region are regulated by specific environmental conditions. Climate reflects continentality, rainfall deficiency, low humidity, hot or warm summers, cold winters and marked temperature fluctuations. As a result little moisture is available for percolation and leaching. Drought is also generally experienced during late summer or autumn months and the surface soils are generally frozen during the winter (Joffe 1936).

In Canada the major areas of grassland soils occupy the cool, subarid to subhumid Interior Plains of the west and minor areas are found in some dry valleys and mountain slopes in the Cordilleran region (Clayton et al. 1977). Plants generally have a short growing season in which to set seed. Spring and early summer temperatures are optimum for grass growth in the steppe flora of these regions whereas the growing cycle is completed by autumn when drying sets in. Elaborate rooting systems develop due to moisture relationships. The fine roots generally die off as the growing cycle is completed and drought sets in. Together with dead stems the fine roots contribute significantly to organic matter and nutrient reserves in the soil.

Grassland soils develop on a great variety of parent materials. The prevailing type on the Canadian scene is glacial drift deposited during the last ice age 10,000-15,000 years B.P. (Legget 1961). The glacial drift generally reflects a variable degree of sorting and is associated with a wide array of landforms. However, one important effect of the ice advance was to incorporate Paleozoic limestone into the drift thereby contributing a calcareous component to the parent material lithology. Calcium and, to a lesser extent, magnesium released through chemical weathering play a major role in base cycling in grassland regions.

It is within the foregoing framework of environmental relationships that features unique to soils of the grasslands develop within the soil body.
Grassland soil formation is a reflection of two major processes: the first and foremost relates to mechanisms of humus formation as part of the biocycle (Fig. 1) and the second, of lesser significance, is the redistribution of bases as part of the hydrological cycle.

Figure 1. The organic matter cycle in the environment.
Growth of mixed herbaceous steppe vegetation during the growing season results in the development of abundant rooting systems that penetrate deep into the soil pedon. While they grow, plant roots release exudates that help build up the mass of organic matter in the soil. Life fades when drought sets in during the late summer and autumn and dead tissue from root systems together with dying leaves and stems add to the reserves of organic matter. Contributions in grasslands can reach 50,000 kg/ha annually, about half of which comes from roots.

Dead organic matter constitutes one of the most widely distributed sources of energy in the soil environment. Plant materials contain a high proportion of cellulose, hemicelluloses and lignins with lesser amounts of other substances such as proteins. But these vary in content with species and age of the plant. Plant materials added to the soil undergo significant changes brought on by lower organisms. But diversity in plant composition as well as environmental conditions influence the nature and direction of humus formation. As an example, the presence of polyphenols can stabilize leaf protein to biological breakdown. Therefore, while it is possible to present a generalized concept for humus synthesis, humus composition and properties reflect the ecosystem in which formation takes place.

Processes of organic matter degradation and humus synthesis are complex in grassland soils but follow a general pathway of ecological activity. Organisms involved include large and small fauna and flora that also contribute their own tissue to the organic matter pool after death. Since soil organisms have rather specific requirements the transformation of organic matter reflects progressive stages in succession, i.e., sequence of organisms occurring in a given time in a given area. Within the soil freshly added organic substrates are colonized by successive waves of microorganisms; each wave altering the micro-habitat and making way for the next group of organisms. The order of succession depends upon a number of factors such as food source changes, amelioration of the environment and suppression by an unfavourable environment.
The initial invasion of organic debris by soil microflora is assisted greatly by activities of soil animals. Animals such as earthworms incorporate surface organic matter into soil, assist in its physical breakdown, and thus expose fresh surfaces to attack. They also ingest organic matter and through action of digestive enzymes produce new forms that remain as fecal deposits. Exoskeletons also contribute constituents to organic matter reserves (Richards 1976).

During degradation there is progressive depletion of chemical energy by heterotrophic organisms. The more simple organic constituents such as sugars, amino acids and organic acids are depleted rapidly since their structures are readily broken down by enzyme systems of zymogenous organisms, providing them with a readily available energy source for their own growth and reproduction. Complex organic compounds such as starches, celluloses, hemi-celluloses, lipids and proteins can also be readily assimilated by microorganisms capable of producing the necessary extra-cellular enzymes required to break them down to simple constituents. Only a few microorganisms are capable of breaking down lignins and lignoproteins so that there is a tendency for these to accumulate in grassland ecosystems. The recalcitrant constituents that accumulate have complex aromatic structures (e.g., polyphenolics) and are comparatively resistant to attack by microorganisms. Therefore, it is through the decomposition of substrates introduced from plants, animals and microbes by zymogenous microorganisms that humus substances build up in the soil.

Chemical changes accompanying organic matter degradation are reflected in narrowing of the carbon : nitrogen ratio. Initially total organic carbon is reduced rapidly during microbial decomposition and respiration returning to the atmosphere in the form of carbon dioxide. Microorganisms assimilate nitrogen that is available from residues or other soil sources to build new protoplasm. The organic carbon : nitrogen ratio falls rapidly to equilibrate at a ratio of 8:1 to 12:1. Most of the nitrogen remains preserved in microbial cells associated with microbial residues and in complex recalcitrant structures. Nitrogen is released gradually from such structures during slow breakdown by autochthonous organisms.
Humus is a combination of relatively stable amorphous humic materials and not a separate chemical entity. It is partially a product of synthesis by various microorganisms involved in its decomposition process and partially recalcitrant remains of original plant substrates; components comprise a high percentage of aromatic structures to which simpler aliphatic groups are attached. As a consequence of its stability, decomposition is extremely slow in grassland soils with mean-residence-times usually ranging from 1000 to 1500 years. This is considered the average time required for the many different organic components to mineralize back to carbon dioxide, water, and inorganic nutrients.

A number of factors can alter the rate of organic matter turnover and hence its mean-residence-time. Cultivation can hasten the process by altering the soil environment. Seasonal additions of readily decomposable organic matter encourages a "flare-up" in zymogenous organisms. The marked increase in microbial activity not only results in rapid breakdown of easily decomposable constituents but has a "priming" effect on the decomposition of recalcitrant constituents and hastens their breakdown as well. After the "flare up" subsides, rate of decomposition returns to a far less but steady level of activity dominated by autochthonous organisms (Gray and Williams 1971). Another important stabilizing effect upon humus is brought about through its interaction with clay minerals in the soil. Humic substances in the presence of 2:1 clay minerals are readily adsorbed by specific adsorption to broken crystal edges and held by strong chemical bonds. The large polymerized humic molecules are physically wrapped around clay particles completely engulfing them. The complexed clay minerals protect the organic matter from decomposition. This stabilization of organic matter is at least partly responsible for the typical granular structures observed in chernozemic A horizons. The mean-residence-time of soil organic matter can also be altered by changes in the nature of the soil organic fractions, soil ecosystem and microclimatic conditions.
The quantity of humus present in the soil reflects the rate of addition and the rate of annihilation (or mean-residence-time). As soil development advances, microbial populations adjust to the availability of food supply, and their level of activity becomes more or less stable (Fig. 2). When additions of organic matter to soil are balanced by losses through mineralization and content remains relatively constant, organic matter turnover attains a "steady-state". The quantity of organic matter present in a steady-state condition is determined by the ratio between the rates of accumulation and annihilation. Either a decrease in the rate at which organic matter is being added to the soil or an increase in the rate of its decomposition can deplete organic matter content and initiate soil degradation.

Figure 2. Development of an Ah horizon as related to time.
Microbial polysaccharides can leach down from the layer of humus accumulation into the B horizon as well where they combine with clays and assist in the development of soil structure (Kononova 1966).

Grassland soils are also characterized by faunal generated soil features that are unique to the environment. Invertebrates make up about 90 percent of the biomass of animals living in the soil. Some of these, such as earthworms and millipedes assist in the development of structure through the production of casts. Other fauna such as ants, termites and small mammals (e.g. moles and gophers) mine from within or from beneath the soil body and serve to mix the soil materials. These processes are manifested in the presence of mounds on the land surface and krotovinas within the pedon. They also have a strong influence upon the character and properties of the soil in grassland landscapes (Hole 1981). As much as 5 percent of area disturbance are noted annually. Up to 50 T/ha of castings from earthworms and 55 T/ha of mixing by moles have been reported annually for some areas.

Mineralization of soil organic matter and biological respiration in the soil markedly increase the partial pressure of CO$_2$ in the soil atmosphere and its solubility in the soil solution. The CO$_2$ charged water initiates the dissolution of otherwise relatively insoluble lime carbonates and enhances their transport down the pedon. At lower depths where the partial pressure of the CO$_2$ is reduced, the soluble bicarbonates revert back to carbonate form and precipitate as a layer of lime carbonate accumulation or calcic horizon.

**EVOLUTION OF GRASSLAND SOILS**

The evolution of grassland soils is based on the assumption that the original colonizers on glacial drift parent material at the time of deposition or shortly thereafter were members of a grassland steppe plant community. In the environments prevailing since that time, climate has been such that redistribution of lime carbonate could occur
without actually being removed from the pedon and remains at a depth where it can be brought well into the pedon during evaporation. Drainage has been sufficient to keep groundwater at a depth where rise of sodium salts to the surface by capillary forces is impossible.

In the early stages of soil development, established grass vegetation contributes an abundance of roots, stems and leaves to develop an organic reserve in the soil. However, initially, there is insufficient decomposition to form an Ah horizon and lime carbonate remains high in the pedon. The soil retains the characteristics of an Orthic Regosol (Fig. 3). The buildup of raw organic matter provides an energy source for an ever increasing soil biological population. Through decomposition, the raw organic matter is transformed into humic constituents. Degradation of organic constituents releases carbon dioxide which enhances the downward movement of lime carbonate through conversion to more soluble bicarbonate form. The buildup of humus and concomitant depletion of lime carbonate at the upper soil surface is initially fairly rapid since large quantities of fresh organic reserves are added seasonally as the grasses ripen during the onset of autumn drought. In the winter months microbial activity slows down. But in the following spring and early summer temperatures reach an optimum and under prevailing moist conditions a flurry of biological activity sets in. Organisms rapidly assimilate the undecomposed organic debris for their growth and reproduction, adjusting their numbers to accommodate the particular ecological conditions in which they exist. Degradation and resynthesis of organic constituents results in formation of humus elements. These too are degraded and mineralized by more persistent autochthonous organisms but at a much reduced rate of activity. At the same time nutrients released by the mineralization of organic matter are mobilized and utilized to stimulate new spring growth in steppe flora. In time an ecological balance is reached, and a relatively constant level of humus content is maintained in the soil, i.e. the soil humus is said to attain a 'steady-state' condition. Under such conditions the microbial population adjusts to the available energy sources which in turn, are controlled by the microenvironment.
Enrichment of humus in the upper pedon is accompanied by strong granular structural development in large part due to the formation of organo-clay complexes but also in part the result of mixing and casting by soil fauna. At this stage of development the surface horizon meets criteria for a chernozemic A and the soil is referred to as Rego Chernozemic subgroup in the Canadian system of soil classification (Canada Soil Survey Committee 1978). As a result of the foregoing, the A horizon has attained its maximum development and is maintained in steady-state.

Continuous liberation of carbon dioxide through organic decomposition and biological respiration stimulates further removal of lime carbonate from the zone immediately below the humus enriched layer and its deposition into the illuvial calcic horizon (Fig. 3). At this point in the pedon, a B horizon forms. Weak hydrolysis and oxidation processes alter iron-bearing minerals and impart a brownish chroma to the soil material. Structural formation is recognized by the presence of a blocky mesostructure and weak prismatic macrostructure. Structural development results from the adsorption of microbial polysaccharides leached from the overlying humic material as well as physical processes associated with seasonal moistening and desiccation, freezing and thawing and root pressure. The development of a Calcareous Chernozemic soil is reflected in the presence of incipient B horizon development, i.e. a Bk. Fully developed Ah and Bm horizons are recognized in sola of Orthic Chernozemic soils. The Orthic subgroup represents a central concept for Chernozemic soils and is considered modal for grassland regions.

However, all Chernozemic soils on our landscapes will not evolve to the Orthic stage. Soil development can attain a steady state condition at any stage in its evolution (Fig. 4). In areas where slopes are steep, infiltration is limited and run-off is prevalent; or where unusually high amounts of lime carbonate are present in the parent material, leaching is generally not sufficient to remove the lime carbonate from the solum, and soils do not develop beyond the Rego or Calcareous Chernozemic stage.
Formation of Soils of the Grasslands

Figure 3. Evolution of grassland soils.

Figure 4. An example of the distribution of Chernozemic soils as related to topographic irregularities in a grassland region.
Landslapes that have attained a steady state condition in their existing environments remain dynamic but soil characteristics and properties exhibit little change. Organic matter additions and removals are in harmony. Nutrients recycle through the root system and are returned to the surface. A relatively uniform rate of turnover keeps the soil cation exchange complex highly saturated with calcium ions, and with lesser amounts of magnesium ions and minor but consistent amounts of potassium and sodium ions. The soil remains near neutral in reaction and microenvironmental conditions favor a wide variety of organisms.

The kind and quantity of humus present is not uniform throughout the grassland regions. Under the True Prairie association (Coupland 1961) within the subarid climatic zone, biomass productivity is significantly restricted by severe moisture deficits during the growing season. Rooting volume is also shallower, thereby contributing to the development of relatively shallow A horizons. The A horizons are thinner (8-12 cm) and have lower organic matter content (3-5%) as compared to the depth (15-30 cm) and organic matter content (10-12%) of their counterparts in the semiarid and subhumid regions. Despite this, there is a relatively rich microflora that rapidly converts the raw organic matter to humus. The accumulated humus has strong brownish chroma as a result of the nature of decomposition. There is less condensation of the aromatic carbon structure with greater number of side-chain radicals in the humic fraction (Fig. 5).

SOILS AND PROCESSES IN THE BOREAL FOREST AND GRASSLAND TRANSITION

An extensive belt lying between the Boreal and Cordilleran forests on the one hand and the grasslands of the Mixed Prairie on the other, described as the Fescue Prairie Grassland (Coupland 1961), has along its northern and western fringes scattered poplar
and poplar–oak stands as part of the plant community. Similar intergrades with scattered
growth of pine, Douglas–fir and poplar form a savannah type parkland in the cooler
subhumid areas of the Palouse Prairie within the intermountain regions (Clayton et al.
1977).

Figure 5. Soil characteristics as related to zonal distribution of climate and vegetation.
It is surmised that the presence of trees reflects three possible historical conditions: (1) forest vegetation slowly encroaching onto grassland as climatic climax vegetation whereas previous fires favored fescues as an edaphic climax vegetation; (2) grasslands establishing on previously forested land through shifts in climate and periodic burning; and (3) the mosaic of forest groves as an adjunct to the Fescue Prairie association, and forest type soils continuing development in that habitat. There is evidence in soil morphology to support all three possibilities.

Under forest cover large amounts of organic matter are added annually to the soil surface as litter fall. Where forest encroaches onto grassland, displacement of organic matter additions from the upper section within the pedon to the landsurface allows for continual depletion of humus by microorganisms without replenishment in the A horizons. The process of degradation results in a characteristic 'salt and pepper' effect, reflecting the presence of bleached zones of low organic content in a darker matrix of higher organic content. Depending upon the manner by which surface leaf litter is incorporated, the rate of degradation may be enhanced through its breakdown. Soil animals readily break down litter and expose it to attack by other organisms. If the litter is acidic and nutrient-poor fungi precede other organisms in succession, their efficient decomposition of leaf litter to forms that are more readily assimilated initiates a 'flare-up' of zymogenous organisms. As the low molecular weight organic constituents leach through the A horizon, the rate of decomposition of recalcitrant components that bind soil aggregates together is enhanced. Released organic acids initiate weathering of less stable minerals and initiate the translocation of disaggregated clays into the B horizon. Hence, under the influence of forest cover the chernozemic horizon is slowly eluviated first to an Ahe and ultimately to an Ae horizon.
Within the forest–grassland transition, grasses frequently follow fire as the prime colonizer of otherwise forested soils. Large contributions of organic matter from a newly established dense rooting system can result in the transformation of eluvial horizons to chernozemic A horizons. Processes are similar to the evolution of Chernozemic soils from freshly exposed parent materials. Either of the foregoing lead to the formation of Dark Gray Luvisolite and Dark Gray Chernozemic soils.

Recently strong evidence has been documented to support the third possibility for genesis of Eluviated Black and Dark Gray Chernozems. Phytolith investigations supported by pollen and historical records suggest much of the forest–grassland fringe area was continuously dominated by poplar cover with a rich understory of shrubs, forbs and grasses. These areas are confined to relatively subdued terrain where moisture regimes are somewhat modified by permanent or temporary water-tables. Initial organic material for humus formation receives contributions from both leaf fall and grasses and is generally substantial since moisture conditions are favourable for high volume growth. The leaf litter is well supplied with bases and is near neutral in pH. These conditions encourage high biological activity and mull formation. Soil-ingesting fauna incorporate the surface leaf litter into the mineral soil where its decomposition by bacteria contributes to the development of stable humus forms. The Dark Gray Chernozems of these areas have L, F and H layers overlying mull A horizons that are difficult to distinguish from degrading and regrading chernozemic A horizons.

**SUMMARY**

Soils and landscapes within the grassland regions reflect the ecosystems through which they evolve and are generally at a steady state in their development.
While natural changes in environmental factors such as climate and vegetation can disrupt the steady state condition in soil development and initiate changes in soil characteristics, such changes are generally gradual and much less profound than those initiated through human activity.

Unregulated, such practices as over-grazing, excessive fallowing, and soil compaction can initiate land degradation through organic matter depletion and accelerated erosion. What has taken nature thousands of years to construct can be destroyed in less than a generation. Yet it is within the capacity of our society not only to protect our lands from abuse, but through proper regulation, develop management plans that will assure their productivity on a sustaining basis.

REFERENCES CITED


DISCUSSION - GRASSLAND SOILS AND PARENT MATERIALS

Specific questions were addressed to the individuals who spoke on the characteristics, dynamics and distribution of Chernozemic soils and surficial materials. The responses are summarized below:

Q. What information is available regarding the rate and dynamics of Ah formation due to invasion by grasses, or conversely degradation as a result of invasion by trees?

A. In the A horizon mean residence time of organic matter gives an indication of the time required to reach a steady state condition, i.e. organic matter breakdown = organic matter addition. Work on Chernozemic soils in Saskatchewan has shown dates for surface organic matter from 750 to 2000 years. When interpreting this type of information it must be remembered that Black Chernozemic soils have 10 to 12% organic matter present in the Ah. However to meet the requirements of a chernozemic A horizon only 1% is required. Therefore, in subhumid conditions with a well-drained soil, the criteria for a chernozemic A horizon might be met in 350 years. However, it would not be in a steady state condition at that point.

Q. How long would the characteristics of a soil profile be retained after a change in vegetation?

A. To answer this question one must consider whether it was originally an edaphic (azonal) or zonal condition which controlled the evolution of the soil. If the intensity of the processes affecting the soil development increase, the previous characteristics of the soil will be obliterated quickly, but, if the intensity decreases, one often finds remnants of the old soil. Salinity, for example, tends to last quite a long time.
Q. How long would you expect the characteristics of a chernozemic A to last after a forest cover becomes established?

A. Degradation begins as soon as the amount of organic matter added to the soil is decreased. In some studies, for example in the Cypress Hills, a recognizable Ac horizon is formed under an aspen stand in 90 years. However, that seems unusually fast.

Q. What are the differences between the clay mineralogy of forested soils and Chernozems which have developed from the same parent material?

A. There can be a differential movement of expansible clay minerals as they eluviate downwards in a forest soil. Expansible clays move downwards first while the coarser, micaceous clays remain behind.

Q. How far does moisture penetrate the Chernozemic profile in the Upper Grasslands?

A. Quantitative information is not available but from direct observations moisture appears to reach into the compact basal till to the depth of the carbonates, which is often at .6 metres. As the season progresses the profile dries out. It seems that coarser soils allow winter moisture to penetrate deeper. Therefore, during the growing season, moisture is available for a longer time than can be expected in a loam textured soil where much of the moisture is retained near the surface and is lost through surface evaporation.
Q. How do the changes with elevation described for the Lac du Bois range grasslands compare with that found in the Cariboo - Chilcotin?

A. In the Lac du Bois area soils change from Brown to Dark Brown to Black Chernozems, with increasing elevation. In the Cariboo - Chilcotin, however, the vertical gradient is less obvious. On low benches beside the Fraser River there are Brown Chernozems and above these Dark Browns. However the plateau soils are predominately Dark Gray Chernozems rather than Black Chernozems. The vegetation zones correspond to those described for the Lac du Bois range, although the Upper Grassland in the Cariboo - Chilcotin is dominated by *Agropyron spicatum* and lacks the *Festuca* species which are common further south.

Q. What management implications are involved with the decomposition products present in the various chernozemic Ah horizons?

A. Little work has been done in this area and what has been reported in the literature has mainly come from Russia. However, it has been shown that in the Brown Chernozemic soils the humic fraction has a lower molecular weight, i.e., the humic acid molecule is smaller, more condensed and more aliphatic. With increasing aridity the type of polymerization that occurs results in a looser, single grain structure which would be a management consideration. This structure is also determined by the amount of bacterial polysaccharides which result from decomposition.

Q. Can Canadian Chernozemic soils be compared to the Chernozems in Russia specifically in regards to the Brown Chernozems?

A. Brown Chernozems in Russia are different from ours because they have predominantly developed in areas which were originally saline and therefore they have solonetzic features (i.e., they are azonal). Historically in the literature, however, they were treated as zonal so it is necessary to keep this in mind when interpreting results.
Q. Are old glacial lake shorelines as distinctive in heavily forested areas as they are in grasslands?

A. In general, probably not, because the steep mountainous topography of most areas likely prevented good glacial lake shoreline formation.

Q. Did the Keewatin (Laurentide) Ice have a substantial influence on the parent material of the Peace River grasslands?

A. One would have to check the available surficial geology maps. However, it is likely that a lot of the tills buried beneath glacial lake sediments in the Peace River area would have been reworked by meltwater streams from the Rocky Mountains or at least from the west. The lake sediments may not show as much mineralogy (parent material) as one might expect from the Laurentide Ice.

Q. In the Peace River area grassland tends to occur on south facing slopes along the river while other slopes support aspen forest. As the topsoil tends to be derived from lake sediments, is there a particular reason for aspen rather than spruce forests?

A. Aspen is likely representative of an early successional stage which is perpetuated by the continual sloughing of the slopes. These slopes are known to be unstable and likely slough-off every one or two hundred or even a thousand years.
POSTER SESSION:
CHARACTERISTICS AND DISTRIBUTION OF CHERNOZEMIC SOILS
IN THE ASHCROFT MAP AREA

M. Fenger

INTRODUCTION

The objective of this paper is to describe the topographic and geographic distribution of the Chernozemic great groups and some selected subgroups within the Ashcroft map area. Chernozemic soils are not extensive within British Columbia; the Ashcroft map area (National Topographic Series 92-I) contains the highest proportion of Chernozems compared to any other map area of similar scale within the province (Valentine et al. 1978). Much of the research on Chernozems has been carried out within this area by the staff of the Agriculture Research Station at Kamloops.

Within the Ashcroft map area there is a general trend, at the great group level of the Canadian System of Soil Classification (Canada Soil Survey Committee 1978), for Chernozems to change from Brown, to Dark Brown, to Black with increasing elevation. Six subgroups within these great groups occur: Orthic Brown (O.B), Rego Brown (R.B), Orthic Dark Brown (O.D.B), Solonetzie Dark Brown (SZ.D.B), Orthic Black (O.B), and Calcareous Black (CA.B).

SOURCES OF DATA AND PROCEDURES

The first major study of grasslands and grassland soils in the Ashcroft area was carried out by Tisdale (1947). In that paper the author recognized three elevational zones; the Lower, Middle and Upper Grasslands, which are characterized by Brown, Dark Brown, and Black Chernozemic soils respectively. A study by van Ryswyk et al. (1966) showed
the relationships between climate, vegetation and grassland soils at different elevations. These studies have provided the basic framework for the Chernozemic portion of the soil legend and the definition of soil associations for soil mapping in the Ashcroft area (Young 1978).

Twenty-three soil associations describe the areas which have dominantly Chernozemic soil profiles (Young 1978). A soil association is a group of relatively homogeneous soils developed on similar parent materials under similar climatic conditions. Soils grouped together in a soil association can be expected to have similar productivity and respond in a similar way to various types of land management practices.

This paper is derived mainly from information contained in the Ashcroft soils report (Young and Fenger in preparation) and is structured according to the zonal framework outlined by Tisdale (1947). Thirty-six Chernozemic profiles were described during the Ashcroft soil survey and are used to summarize the characteristics of the great groups for this area (B.C. Min. Environ. unpublished data*).

Grassland soils and associated grassland vegetation have been the subject of several other studies in the Ashcroft map area including those of Spilsbury and Tisdale (1944), Weir (1955), McLean and Marchand (1968), Lord and McLean (1969), Watson (1977), and Jakoy (1981).

CHERNOZEMS WITHIN THE LANDSCAPE

The diagnostic criterion for a Chernozemic soil is the presence of an A horizon, in

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*Soil profiles from within the Ashcroft map area are stored within the B.C. Soil Information System file and are available on request from the Survey and Resource Mapping Branch, British Columbia Ministry of Environment, Victoria, B.C.
which organic matter has accumulated. The requirements of a chernozemic A horizon are defined by the Canada Soil Survey Committee (1978) and summarized by Green and van Ryswyk (this publication).

Figure 1 shows the general elevational distribution of taxonomic soil orders and great groups in the map area, and Figure 2 shows the geographic distribution of the Chernozemic subgroups. The Orthic subgroups of the Brown, Dark Brown and Black great groups are the most extensive and occur as the dominant or modal soil for most soil associations. The Rego, Solonetzic and Calcareous subgroups are less extensive; each are dominant in one soil association.

THE BROWN CHERNOZEMIC GREAT GROUP

Brown Chernozems are differentiated from other great groups by colour, as their name implies. They must have a chernozemic A which has a Munsell (1975) colour value darker than 3.5 moist and 4.5 to 5.5 dry and a chroma which is usually darker than 1.5 (Canada Soil Survey Committee 1978). The Brown Chernozemic soil climate is typically cold and subarid (Clayton et al. 1977). These soils are restricted to the lower elevations along the major valleys of the Thompson and Fraser rivers. Associated with these soils, Tisdale (1947) delineated a "Lower Grassland Zone" characterized by Agropyron spicatum and Artemisia tridentata. This zone was also recognized by van Ryswyk et al. (1966) as the Artemisia tridentata zone.

The profile development of the Brown Chernozems within the map area typically consists of a thin A horizon which is often restricted to a fine capping of aeolian materials. The nine profiles sampled within this great group were between 335 and 525 m in elevation with an average of 420 m. The A horizons of the profiles ranged from 10 to 30 cm in depth with an average of 17 cm, and organic carbon content ranged from 1.0 to
Note that there is overlap between soil orders and that GLEYSOILS, ORGANICS and REGOSOLS do not follow this pattern and can occur at any elevation.

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<thead>
<tr>
<th>Biogeoclimatic Zones</th>
<th>Biophysical Forest Zones and Subzones</th>
</tr>
</thead>
<tbody>
<tr>
<td>Engelmann spruce - Subalpine fir Zone</td>
<td>Subalpine Engelmann spruce - Alpine fir Zone</td>
</tr>
<tr>
<td>Interior Douglas-fir Zone and Montane spruce Zone</td>
<td>Interior Douglas-fir Zone Lodgepole pine subzone</td>
</tr>
<tr>
<td>Ponderosa pine Bunchgrass Zone</td>
<td>Interior Douglas-fir Zone ponderosa pine subzone</td>
</tr>
<tr>
<td>Interior Bunchgrass Zone</td>
<td></td>
</tr>
</tbody>
</table>

Figure 1. Schematic diagram showing the general trend in soil great groups with increasing elevation within the Ashcroft map area.

1 after Krajina (1960) and Mitchell and Green (1981)

2 after van Bornveld (1977), described in Young and Fenner (in preparation)
<table>
<thead>
<tr>
<th>Map Symbol</th>
<th>Subgroup</th>
<th>Great Group</th>
<th>Soil Association</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Rego</td>
<td>Brown</td>
<td>Cache Creek</td>
</tr>
<tr>
<td></td>
<td>Orthic</td>
<td>Brown</td>
<td>Carabine</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Courtenay</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Flat Creek</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Godey</td>
</tr>
<tr>
<td></td>
<td>Solonetzi</td>
<td>Dark Brown</td>
<td>McKnight</td>
</tr>
<tr>
<td></td>
<td>Orthic</td>
<td>Dark Brown</td>
<td>Tranquille</td>
</tr>
<tr>
<td></td>
<td>Orthic</td>
<td>Black</td>
<td>Lac du Bois</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Souss</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Glimpse</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>McQueen</td>
</tr>
<tr>
<td></td>
<td>Calcareous</td>
<td>Black</td>
<td>Commonage</td>
</tr>
<tr>
<td></td>
<td>Orthic</td>
<td>Black</td>
<td>Laluwissen</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Trapp Lake</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Medicine</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Alymer</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Mossey</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Gwenn</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Trachyte</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Meander</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>Tulle</td>
</tr>
</tbody>
</table>

Figure 2. Soils legend for distribution of Chernozemic subgroups of the Ashcroft Map Sheet, 92-l.
1.9% with an average of 1.3% (B.C. Min. Environ. unpublished data). B horizons were also thin; carbonates had only leached to a depth of 10 to 30 cm before precipitating to form a carbonate enriched horizon.

Figure 3 shows the nine soil associations dominated by soils of the Brown Chernozemic great group. The subgroup classification, parent materials and dominant bedrock types pertaining to these soil associations are also shown. Table 1 summarizes the texture, drainage, and geographic distribution of the five most extensive soil associations.

Table 1. Characteristics and distribution of the most extensive soil associations (of the Ashcroft map area [92-1]) with dominantly Brown Chernozemic development

<table>
<thead>
<tr>
<th>ASSOCIATION</th>
<th>TEXTURE</th>
<th>DRAINAGE</th>
<th>DISTRIBUTION</th>
</tr>
</thead>
<tbody>
<tr>
<td>Courtenay</td>
<td>sandy loam to loam</td>
<td>well</td>
<td>Thompson River between Savona and Spences Bridge</td>
</tr>
<tr>
<td>McKnight</td>
<td>loam to silty loam</td>
<td>well</td>
<td>near Ashcroft and south towards Spences Bridge, south aspects along South Thompson River and north of Kamloops</td>
</tr>
<tr>
<td>Godey</td>
<td>sandy loam to loam</td>
<td>rapid</td>
<td>Lytton to Lillooet and Ashcroft to Kamloops</td>
</tr>
<tr>
<td>Lundbom</td>
<td>silty clay loam to clay loam</td>
<td>well</td>
<td>along South Thompson River</td>
</tr>
<tr>
<td>Cache Creek</td>
<td>sandy loam to loam</td>
<td>well</td>
<td>Semlin and Bonaparte valleys</td>
</tr>
</tbody>
</table>
Figure 3. Brown Chernozemic soil associations of the Ashcroft map area (92-I).
THE DARK BROWN CHERNOZEMIC GREAT GROUP

Dark Brown Chernozems have an A horizon somewhat darker than the Brown great group with Munsell (1975) colour values darker than 3.5 moist and 3.5 to 4.5 dry and a chroma usually greater than 1.5 dry (Canada Soil Survey Committee 1978). The Dark Brown soil climate is typically cold and semi-arid (Clayton et al. 1977). Associated with the area of Dark Brown Chernozemic soils Tisdale (1947) classified a "Middle Grassland Zone" characterized by Agropyron spicatum and Poa sandbergii. This zone was also recognized by van Ryswyk et al. (1966) but described by a Stipe comata and Poa sandbergii grassland community.

The Dark Brown Chernozemic soil profile is typically thicker and better expressed than profiles described within the Brown great group. The nine profiles sampled within this great group ranged in elevation from 350 to 930 m with an average of 600 m; A horizons ranged from 10 to 34 cm with an average of 19 cm and organic carbon content ranged from 1.2 to 3.8% with an average of 2.3% (B.C. Min. Environ. unpublished data).

Figure 4 shows the seven soil associations dominated by soils of the Dark Brown Chernozemic great group; the subgroup classification, parent materials and dominant bedrock types are also illustrated. Table 2 summarizes the texture, drainage and geographic distribution of the most extensive soil associations shown in Figure 4. The Dark Brown great group occurs intermittently above the Brown great group along the Thompson River and south of Kamloops to Nicola Lake at the lowest elevation of the Thompson Plateau (Figure 2).
Figure 4. Dark Brown Chernozem soil associations of the Ashcroft map area (92-D).
Table 2. Characteristics and distribution of the most extensive soil associations with dominantly Dark Brown Chernozemic development

<table>
<thead>
<tr>
<th>ASSOCIATION</th>
<th>TEXTURE</th>
<th>DRAINAGE</th>
<th>DISTRIBUTION</th>
</tr>
</thead>
<tbody>
<tr>
<td>Glimpse</td>
<td>sandy to loamy sand</td>
<td>rapid</td>
<td>extensive; concentrations in area bounded by Knutsford, Trapp and Napier lakes, also in area bounded by Nicola, Douglas and Minnie lakes.</td>
</tr>
<tr>
<td>Lac du Bois (weakly saline)</td>
<td>silty clay loam to loam</td>
<td>well</td>
<td>limited distribution, near Merritt and Nicola Lake</td>
</tr>
<tr>
<td>McQueen</td>
<td>silty loam to silty clay loam</td>
<td>well</td>
<td>north and south of Kamloops and north of Ashcroft</td>
</tr>
<tr>
<td>Trapp Lake</td>
<td>silty loam to silty clay loam</td>
<td>well</td>
<td>scattered units within an area bounded by Ashcroft, Savona, Knutsford and Douglas Lake</td>
</tr>
</tbody>
</table>

THE BLACK CHERNOZEMIC GREAT GROUP

The A horizon of the Black Chernozems is the darkest of the three great groups discussed, with Munsell (1975) colour values darker than 3.5 dry and chroma usually 1.5 or less moist (Canada Soil Survey Committee 1978). The Black Chernozemic soil climate is typically cold and subhumid (Clayton et al. 1977). Black Chernozemic soils are most extensive east of Merritt and are situated away from the major river valleys on the undulating surface of the Thompson Plateau.

Associated with the area of Black Chernozemic soils Tisdale (1947) delineated an "Upper Grassland Zone" dominated by *Agropyron spicatum* and *Festuca scabrella*. This zone, characterized by *Festuca scabrella*, was also recognized by van Ryswyk et al.
(1966) who further subdivided it into an upper and lower portion based on differences in elevation, plant species present, April to October mean temperature and the amount of organic carbon present within the upper 2 dm of the solum.

The Black Chernozemic soil profiles have the thickest and most well developed A horizons. The 18 profiles sampled ranged in elevation from 880 to 1220 m with an average of 1050 m; A horizons ranged from 10 to 35 cm in thickness with an average of 21 cm and organic carbon content ranged from 2.1 to 5.7% with an average of 3.2% (B.C. Min. Environ. unpublished data).

Figure 5 shows the seven dominant Black Chernozemic soil associations and the subgroup classification, parent materials and dominant bedrock types pertaining to them. Table 3 summarizes the texture, drainage, and geographic distribution of the most extensive soil associations shown in Figure 5.

Table 3. Characteristics and distribution of the most extensive soil associations with dominantly Black Chernozemic development

<table>
<thead>
<tr>
<th>ASSOCIATION</th>
<th>TEXTURE</th>
<th>DRAINAGE</th>
<th>DISTRIBUTION</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mossey</td>
<td>loam to silty loam</td>
<td>well</td>
<td>North Thompson River Valley south of Kamloops and west of Cache Creek</td>
</tr>
<tr>
<td>Trachyte</td>
<td>silty loam to silty clay loam</td>
<td>well</td>
<td>east and south of Nicola Lake</td>
</tr>
<tr>
<td>Tullee</td>
<td>silty loam to silty clay loam</td>
<td>well</td>
<td>near Douglas and Minnie lakes</td>
</tr>
</tbody>
</table>
Figure 5. Black Chernozemic soil associations of the Ashcroft map area, (92-I).

SUMMARY

The Ashcroft map area (National Topographic Series 92-I) contains the highest proportion of Chernozemic soils of any map area within British Columbia. Within this
area there is a general trend, at the great group level of the Canadian System of Soil Classification, for Chernozems to change from Brown, to Dark Brown, to Black with increasing elevation. This corresponds to an increase in the thickness and amount of organic carbon within the A horizons with increasing elevation.

Below the great group level six subgroups occur: Orthic Brown, Rego Brown, Orthic Dark Brown, Solonetzic Dark Brown, Orthic Black and Calcareous Black. Descriptions of the characteristics and distribution of the Chernozemic soil associations of the Ashcroft Map area provided here can assist in extrapolation of research findings leading to improved understanding and management of grasslands.

REFERENCES CITED


