44679

National Library of Canada
Canadian Theses Division
Ottawa, Canada
K1A 0N4

Bibliothèque nationale du Canada
Division des thèses canadiennes

PERMISSION TO MICROFILM — AUTORISATION DE MICROFILMER

- Please print or type — Écrire en lettres moulées ou dactylographier

Full Name of Author — Nom complet de l'auteur
ABDER-RUHMAN, MOHAMMED

Date of Birth — Date de naissance
1st Feb. 1949

Country of Birth — Lieu de naissance
Palestine

Permanent Address — Résidence fixe
Burqa- Nablus
WEST BANK PALESTINE

Title of Thesis — Titre de la thèse
Mineralogical Characteristics Of Soils from East Central Alberta

University — Université
The University Of Alberta

Degree for which thesis was presented — Grade pour lequel cette thèse fut présentée
M.Sc

Year this degree conferred — Année d'obtention de ce grade
1980

Name of Supervisor — Nom du directeur de thèse
Dudas, M.J

Permission is hereby granted to the NATIONAL LIBRARY OF CANADA to microfilm this thesis and to lend or sell copies of the film.

The author reserves other publication rights, and neither the thesis nor extensive extracts from it may be printed or otherwise reproduced without the author's written permission.

L'autorisation est, par la présente, accordée à la BIBLIOTHÈQUE NATIONALE DU CANADA de microfilmer cette thèse et de prêter ou de vendre des exemplaires du film.

L'auteur se réserve les autres droits de publication; ni la thèse ni de longs extraits de celle-ci ne doivent être imprimés ou autrement reproduits sans l'autorisation écrite de l'auteur.

Date
24th Dec. 79

Signature
M.A.

NL-91 (4/77)

National Library of Canada
Collections Development Branch

Canadian Theses on
Microfiche Service

Bibliothèque nationale du Canada
Direction du développement des collections

Service des thèses canadiennes
sur microfiche

NOTICE

The quality of this microfiche is heavily dependent upon the quality of the original thesis submitted for microfilming. Every effort has been made to ensure the highest quality of reproduction possible.

If pages are missing, contact the university which granted the degree.

Some pages may have indistinct print especially if the original pages were typed with a poor typewriter ribbon or if the university sent us a poor photocopy.

Previously copyrighted materials (journal articles, published tests, etc.) are not filmed.

Reproduction in full or in part of this film is governed by the Canadian Copyright Act, R.S.C. 1970, c. C-30. Please read the authorization forms which accompany this thesis.

THIS DISSERTATION
HAS BEEN MICROFILMED
EXACTLY AS RECEIVED

AVIS

La qualité de cette microfiche dépend grandement de la qualité de la thèse soumise au microfilmage. Nous avons tout fait pour assurer une qualité supérieure de reproduction.

S'il manque des pages, veuillez communiquer avec l'université qui a conféré le grade.

La qualité d'impression de certaines pages peut laisser à désirer, surtout si les pages originales ont été dactylographiées à l'aide d'un ruban usé ou si l'université nous a fait parvenir une photocopie de mauvaise qualité.

Les documents qui font déjà l'objet d'un droit d'auteur (articles de revue, examens publiés, etc.) ne sont pas microfilmés.

La reproduction, même partielle, de ce microfilm est soumise à la Loi canadienne sur le droit d'auteur, SRC 1970, c. C-30. Veuillez prendre connaissance des formules d'autorisation qui accompagnent cette thèse.

LA THÈSE A ÉTÉ
MICROFILMÉE TELLE QUE
NOUS L'AVONS REÇUE

Ottawa, Canada
K1A 0N4

NL-339 (Rev. 8/80)

THE UNIVERSITY OF ALBERTA

MINERALOGICAL CHARACTERISTICS OF SOILS FROM EAST CENTRAL ALBERTA

by

MOHAMMED ABDER-RUHMAN

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH

IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE

OF MASTER OF SCIENCE

IN

SOIL SCIENCE

EDMONTON, ALBERTA

SPRING

1980

THE UNIVERSITY OF ALBERTA

FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled MINERALOGICAL CHARACTERISTICS OF SOILS FROM EAST CENTRAL ALBERTA submitted by MOHAMMED ABDER-RUHMAN in partial fulfilment of the requirements for the degree of MASTER OF SCIENCE in SOIL SCIENCE.

M. J. Dudas

Supervisor

Gerald Coen

S. Pawluk

Date December 7, 1979

To
My Parents
My Wife Khitam
My Children Reema, Anas and Beesan

ABSTRACT

Five pedons representing common soils in the Cold Lake-Bonnyville area were investigated to fully characterize their mineralogical attributes. The soils consisted of two Orthic Gray Luvisols, two Dark Gray Luvisols and an Eluviated Black Chernozemic. One additional pedon of an Orthic Gray Luvisol from the Cooking Lake moraine near Edmonton was also included in this study. Except for minor variations, types and nature of minerals present in all six soils were generally similar.

Soils located further from the Shield area contained lower amounts of feldspars and mica than those in closer proximity to the Shield. The common abundant phyllosilicates in the clay fraction of all soils were smectites, kaolinite and mica. Vermiculite constituted less than 15 per cent of the clay fraction while chlorite was present in minor amounts. Distribution of clay minerals within profiles indicated smectite was preferentially translocated downward. Distribution of vermiculite was interpreted to reflect an inherited feature of parent materials. Dioctahedral clay-sized mica displayed a relatively uniform distribution within all profiles. Quartz and minor amounts of feldspars were present in the clay fraction from all soil horizons. Accumulation of primary clay-sized quartz in eluviated horizons was particularly evident.

Sand and silt fractions were dominated by primary minerals. Quartz was the dominant mineral species in sand and silt fractions accompanied by smaller amounts of plagioclase, potassium feldspars and heavy minerals (biotite, amphiboles, vermiculite and chlorite). Kaolinite was identified in fine silt fractions and talc was present in the silt fraction. Contents of plagioclase feldspars in soils were generally one to two times that of potassium feldspars. Plagioclase feldspars were dominated by albite (or albitic feldspar) with small amounts of oligoclase and other calcic feldspars. Potassium feldspars consisted of orthoclase, microcline and perthites which displayed a high degree of exsolution.

Very little weathering was observed for sand-sized potassium feldspars while somewhat pronounced weathering was observed for plagioclase feldspars. Trioctahedral mica in silt fractions appeared weathered to vermiculite in surface horizons when compared to C horizons. Moderate weathering of amphiboles in both sand and silt fractions occurred in response to pedogenesis; the absence of clay-sized amphiboles indicated their dissolution when comminuted to clay-size particles. In general, pedogenic development at the six sites was comparable in its effects on the mineralogical attributes of the soils. The main effects of pedogenesis were: (1) redistribution of the major mineral constituents of the clay fraction (especially smectite) in response to clay translocation and (2) some weathering of feldspars (particularly plagioclase feldspars), biotite, amphiboles and chlorite in sand and silt fractions from the upper solum.

Parent material uniformity was examined to assist in the interpretation of observed variations in mineral constituents in terms of either weathering or inherited characteristics. Parent materials at four of the sites appeared uniform while wind and/or water action apparently modified surface horizons of the other two sites. Detailed particle size data proved to be superior in identifying major parent material discontinuities and/or small modifications such as eolian additions. Size distribution of resistant minerals was also equally sensitive as a measure of parent material uniformity.

ACKNOWLEDGEMENTS

Deep and sincere appreciation is extended to Dr. M.J. Dudas for his kind guidance, full encouragement and constructive criticism during the course of this study and for his assistance and unlimited patience in the preparation and reviewing of the manuscript.

Further thanks are extended to Dr. S. Pawluk, for his encouragement during the course of this study. Special appreciation is also expressed to Dr. R. May, Associate Professor of Geology, for his guidance in statistical procedures and to Dr. G.M. Coen for serving on the examining committee. Special and sincere thanks are also expressed to the Department of Soil Science staff and graduate students for the unforgetable friendly and scientific atmosphere.

Acknowledgements are also expressed to Dr. J. Robertson, Mr. G. Spiers, Mr. B. Howitt, Mr. S. Hmeidan, Mr. A. Abu-Ulba, Mr. P. yee and Mrs. S. Putz for their valuable comments and frequent valuable assistance to the author; Ms. Louiza Gordon for typing the manuscript.

The author is indebted to the Alberta Institute of Pedology and the Soil Science Department for financial assistance made available for this investigation.

The author would like to express his deep and sincere appreciation to his wife, Khitam, for her encouragement and many sacrifices during the course of the study. Further deep and sincere appreciation are extended to all my family in Palestine for their continuous encouragement and financial assistance.

Table of Contents

Chapter Page

LIST OF TABLES

Table Page

LIST OF FIGURES

LIST OF PLATES

Plates Page

LIST OF APPENDIX TABLES

Table Page

LIST OF APPENDIX FIGURES

I. INTRODUCTION

Only limited information exists on the mineralogical nature of soils in Western Canada despite the fact that decisions on the use and management of these soils are particularly dependent on knowledge of the properties of the inorganic fraction. Documentation of the mineralogical characteristics of soils in Alberta has largely been confined to a few pedological studies where the major emphasis was not the accurate qualitative or quantitative evaluation of soil mineral components. In some studies on Alberta soils, different investigators examining the same soil report very dissimilar results for their quantitative soil mineralogical assessments particularly for the clay mineral fraction. Such discrepancies further point out the need for detailed and accurate soil mineralogical studies.

The present study was initiated to obtain comprehensive information on the mineralogical nature of selected soils in the Cold Lake and Bonnyville area of east central Alberta. Soils from this area were chosen for examination because no information is available on their mineralogical attributes and it was felt that soils from this eastern part of the province, being in close proximity to the Canadian Shield, might display mineralogical characteristics quite different than those of soils previously examined in more western and southern regions of Alberta. Only a limited number of Chernozemic and Luvisolic soils were included in this study in order that major effort could be placed on detailed quantitative and qualitative mineralogical investigations rather than partial characterization of numerous samples. Soils collected for this study were examined with the following objectives in mind:

1. to qualitatively and quantitatively evaluate the nature of the clay mineral fraction
2. to characterize the mineralogical properties of the sand and silt-sized fractions with major emphasis on feldspar mineralogy
3. to assess the extent and nature of mineral alteration accompanying pedogenesis.

To accomplish the third objective it was necessary to evaluate parent material uniformity using established methods common to pedological research.

II. LITERATURE REVIEW

A. Mineralogical Composition Of Soils Of The Interior Plains

The nature of surficial deposits and their origin is an important aspect of the mineralogical characteristics of the soils of the Interior Plains. Therefore, the geological history of the Plains will be briefly discussed in this section.

The Interior Plains are the largest physiographic area to the west and southwest of the Canadian Shield (Bostock,1970). In Canada, it extends from the 49th parallel to the Mackenzie Delta between the Canadian Shield and the mountains of the Cordilleran region. The Plains are underlain by gently dipping sedimentary rock of Paleozoic, Cretaceous and Tertiary ages. The Paleozoic rocks underlie a narrow strip of the Interior Plains adjacent to the Canadian Shield; these sedimentary rocks consist of limestone and dolomite of Ordovician, Silurian and Devonian age. In Manitoba, Jurassic shale bedrock is present to the west of the Paleozoic rocks. Excluding Paleozoic and Jurassic rocks, the rest of the Interior Plains are mainly underlain by Cretaceous sedimentary rocks (Bostock, 1970). In Alberta, most of the northern, east-central and southern Plains are underlain by interbedded marine and nonmarine Cretaceous sandstones and silty shales with subordinate amounts of coal, bentonite clay-ironstone and volcanic ash. The west-central portion along with regions adjacent to the foothills are underlain by nonmarine sandstones and shales of Cretaceous and early Tertiary age (Pawluk and Bayrock, 1969).

Erosional processes during Tertiary times associated with easterly flowing preglacial drainage systems shaped the major elements of the physiography of the Interior Plains. The major physiographic subdivisions of the plains consists of Tertiary erosional remnants including the Turtle Mountains in Manitoba, Wood and Moose Mountains in Saskatchewan, the Cypress Hills of Saskatchewan and Alberta, Hand Hills, Neutral Hills, Swan Hills and Caribou and Horn Mountains of Alberta. The southern portion of the Interior Plains is broken into three levels or steppes by the Manitoba Escarpment and the Missouri Coteau which gives rise to the Manitoba, Saskatchewan and Alberta Plains. The later rises 60-150 meters above the Saskatchewan Plain and up to 1250-1400 meters a.s.l near the foothills in the west

The physiography of the Interior Plains, though shaped during the Teritary period, was altered through erosional and depositional processes associated with glaciation during the Quaternary period. Repeated cold periods during the Pleistocene resulted in ice sheets periodically covering virtually all of the Canadian landscape during the last million years. Wisconsin glaciation was an extremely complex event, but in simple terms it comprised of repeated advances and retreats of the ice with interglacial periods of warm temperatures. The last Wisconsin glaciation consisted of two major ice sheets. The Cordilleran ice sheet buried the entire Western Cordillera from the Coast Range to the Rockies (Bird, 1972). The second or Laurentide ice sheet covered an area extending from the Rocky Mountains to the Continental Shelf southwest of Nova Scotia (Bird, 1972). The slow and radial outward movement of the ice from its Keewatian center was generally in a south-southwest direction across the southern Plains and the Alberta Plateau to meet with the Cordilleran ice sheet of the Rockies and then flowed in a southeasterly direction roughly parallel to the foothills region in southern Alberta (Gravenor and Bayrock, 1965). The direction of the ice advance from the Canadian Sheild to the northwestern section of the Plains was to the north and northwest (Clayton et al., 1977). Incorporation of granitic rock material transported from the Precambrian Shield as well as limestone and dolomite from the adjacent Paleozoic beds with the underlying bedrock and the unloading of these materials accompanied by reworking by water and wind following the ice retreat resulted in the formation of landform features recognized as hummocky, ridged, fluted, rolling and undulating morainal plains, outwash and lake plains, drumlins, eskers, kettles and glaciofuvial plains.

Surficial deposits of glacial origin are the main parent materials of the soils developed in the Interior

Plains. Some soils have developed on preglacial unconsolidated deposits in the non-glaciated areas including the top of the Cypress Hills and portions of the Wood Mountain Plateau (Bird, 1972). Glacial deposits occuring throughout the Interior Plains include till which is unsorted structureless material deposited directly from the ice and stratified drift which is a sorted and layered meltwater sediment deposited along the margin of ablating glacial ice. Other glacial deposits include lacustrine deposits comprised of silts and clays. These deposits formed in glacial lakes which developed after the ice retreat due to blocking of preglacial valleys or from deformation of the earth's crust by the weight of the ice (Bird, 1972). Enormous areas were covered with glacial lakes throughout the Interior Plains. Major glacial lakes include the Edmonton, Red Deer, Rycroft, Fahler and Peace lakes in Alberta, the Swift Current, Rosetown, Saskatoon and Regina lakes in Saskatchewan and Lake Agassiz in southern Manitoba (Bird, 1972). Deposits resulting from the reworking of glacial materials by either river waters or winds in the postglacial period include glaciofluvial sediments in the areas adjacent to river valleys and their deltas and eolian deposits scattered in some areas throughout the Plains region and more extensively in southern Saskatchewan and Alberta.

Information on the geological origin of surficial deposits, though difficult to ascertain, was demonstrated to be valuable in the interpretation of the composition of till deposits in Alberta (Pawluk and Bayrock, 1969; Twardy et al., 1974). Since most of the soils in the Interior Plains are short term features on the geological scale, one would expect that the major mineralogical attributes of these soils are inherited from their parent material and not due to authigenesis accompanying soil formation. The possible origin of most soil parent material in the Interior Plains is the underlying bedrock material with varying amounts of Precambrian or Paleozoic rock transported from the Shield and adjacent Paleozoic strata. Establishing a strong relationship between till and underlying bedrock is complicated by the fact that till may be the result of repeated glacial advance and retreats from different directions. However, ice contact deposits do tend to reflect the local bedrock lithology as the mode of deposition appears to have been rapid in most cases with a limited distance of transport (Flint, 1971). Other glacial or glacial derived deposits such as glacio-fluvial, glacio-lacustrine and eolian deposits do not usually reflect only the underlying bedrock since materials constituting these deposits can be transported from distant locations with possible admixture of materials from several sources. The following factors increase the complexity and reduce the reliability of the estimates of local bedrock contribution to the lithology of the glacial deposits: (1) the uncertainity of the direction of ice movement, (2) repeated nature of the glaciation process, (3) possible changes in the ice movement from one glacial advance and retreat to another and (4) dependance of the types and amounts of materials incorporated with the ice load on the kinds and topography of the bedrock traversed by the ice. These factors are difficult to resolve, however, consideration of the lithology of both regional and local bedrock together with knowledge of the established ice directions would help in the evaluation of bedrock source material for glacial deposits.

The mineralogy of the major bedrock materials in the Interior Plains was studied in northeastern, east and west-central as well as southern Alberta. Paleozoic and Mesozoic rocks in northeastern Alberta were studied by Carrigy (1959). He reported Paleozoic rocks from this region to consist of evaporites including calcite, dolomite, halite, gypsum, anhydrite and organic carbonates. The range in content of the dolomitic limestone was 61 to 94 per cent calcium carbonate and 34 to 1 per cent magnesium carbonate with minor amounts of silica, aluminium and iron oxides. The Mesozoic rocks underlying parts of the Fort McMurray area were reported to consist of sands, sandstones, siltstones and shales. The dominant clay minerals in the shales were illite and kaolinite while the siltstones consisted of abundant amounts of quartz with minor amounts of kaolinite, illite and chlorite. The heavy mineral assemblage of the sands indicate their derivation from Precambrian Shield rocks and were reported to contain tourmaline, chlorite, zircon, garnet, kyanite, apatite and rutile while quartz and small amounts of orthoclose constituted the bulk of the sands. The mineralogical composition of the Tertiary and Cretaceous sandstones underlying the central and southern Alberta Plains was also reported by Carrigy (1971). His results indicate that these sandstones consist of variable amounts of quartz, chert, volcanic and non-volcanic rock fragments, feldspars, calcite and dolomite. The average proportion of feldspars and calcite in the sandstone of the Porcupine Hill Formation was 3% and 9%, respectively, while that of the Paskapoo Formation was 10% and 3%, respectively. The latter Formation contained less quartz and chert and more volcanic fragments. Furthermore, he indicated that upper Paskapoo sandstones appear to have a higher proportion of potash feldspar than the lower Paskapoo sandstones in west-central Alberta. The heavy mineral assemblage of the Porcupine Hills Formation consisted of abraded grains of zircon, tourmaline and apatite while the Paskapoo Formation was dominated by biotite, zircon, apatite, epidote and hornblende. The cementing agents present in the sandstones consist of calcite, dolomite and clay minerals including kaolinite, montmorillonite and chlorite. Earlier, Mellon (Pawluk and Bayrock, 1969) reported on the content of feldspars in some Cretaceous sandstones of different formations in northern Alberta. His results indicated the content of sodium-calcium feldspars ranged from 1 to 23 %

while content of potassium feldspars ranged from 0.6 to 11 % . The fine grained portions of the near surface bedrocks of central Alberta were studied by Locker (1973) who found them to display a wide variation in composition. In general, siltstones and silty-claystones consisted of fine sand and silt size particles of quartz, feldspars, rock fragments and mica admixed with variable proportions of clay minerals, organic matter, carbonates and pyrite. Semi-quantitative estimates of clay minerals from X-ray diffraction analysis revealed that montmorillonite was the dominant clay in the formations underlying the eastern parts of the Central Plains region while illite (hydrous-mica) was the most common clay mineral in the western part of the region. The average montmorillonite content of the fine-grained portion of the Belly River and Lower and Upper Edmonton Formations underlying the east-central region of Alberta was over 50% while that of the Paskapoo Formation was less than 50% with comparable but reverse values for content of illite. However, the average contribution of montmorillonite, illite and kaolinite plus chlorite to the bulk composition of the rock was 23, 10 and 5%, respectively, for the Belly River Formation and 9, 12 and 5%, respectively, for the Paskapoo Formation. Mineralogical composition of the clay fraction in shales of the Bearpaw Formation in southern Alberta was investigated in three sections by Byrne and Farvolden (1959). Their results indicated a predominance of montmorillonite and illite and the presence of chlorite rather than kaolinite as a third major constitutent. The mineral constituents of the upper portion of the Formation based on X-ray diffraction peak intensities ranged from 61±17% to 87±10% for montmorillonite, 9±7% to 25±12% for illite and 4±3% to 14±5% for chlorite. Other minerals present in the clay fraction were quartz; cristobalite and feldspars with little evidence for the presence of mixed layered clays. The broad nature of diffraction peaks suggested some of the minerals were very fine-grained or had defects in their crystal structure or both. The results of the Greene-Kelley test involving X-ray diffraction of lithium treated samples and heated to 200°C indicated the presence of octahedral substitution in the smectites([illegible]g., beidellitic montmorillonte). Similar contents of clay mineral constituents in shales and bentonitic shale layers of the Bearpaw Formation were reported by Forman and Rice (1959) for drill holes in southern Saskatchewan. The composition of the fine, medium and coarse silts were predominantly quartz with the exception of one sample which was predominated by illite. Other major minerals present in the silt fraction included illite and feldspars in the coarse and medium silt with kaolinite mainly present in the fine silt. Sand fractions (>50 micrometer) displayed some variations in mineralogy with the >3.0 g/cc specific gravity separate predominated by pyrite or biotite. The 3.0-2.7 g/cc specific gravity separate was predominated by calcic feldspar or biotite and the <2.7 g/cc specific gravity separate predominated by quartz. Other minerals present included chlorite, kaolinite, opaques, pyroxenes and zircon.

On the basis of the foregoing discussion and the limited detailed quantitative information available on mineral composition of bedrock materials in the Interior Plains region, it is difficult to delineate or recognize discrete mineralogical regimes associated with different bedrock formations. A broad and recognizable feature is that montmorillonite is the dominant phyllosilicate in the clay sized fraction in bedrock materials underlying the central and southern portion of the Interior plains of Alberta with a dominance of illite in bedrock located in the western and foothills regions. Kaolinite appears to be the dominant phyllosilicate in the clay fraction from bedrock in the northeastern part of Alberta. Other minerals including chlorite, quartz, calcite, dolomite and feldspars usually occur with montmorillonite, illite and kaolinite in the sedimentary bedrock formations. No systematic regional trends are evident for primary mineral composition of sedimentary bedrock in Alberta.

A limited number of studies have been published on the mineralogical composition of surficial materials in the Interior Plains region. The mineralogical, chemical and physical characteristics of several till samples from Alberta were documented by Pawluk and Bayrock (1969). These investigators found the <2 micrometer clay fraction of till to contain variable amounts of montmorillonite, illite, kaolinite and quartz. Content of montmorillonite varied from 20 to 60% with the highest quantities found in central and southern regions and the lowest in the northern regions where hydrous mica content was high (35-48%). Interlayering of montmorillonite with hydrous mica was evident and appeared to be confined largely to the tills of southern and central Alberta. Dioctahedral montmorillonite having a composition between that of beidellite and nontronite was the type of smectite reported to be present. Kaolinite and quartz occurred in relatively low quantities. There was little or no evidence for the presence of feldspars in the clay fractions. Some samples were reported to contain substantial amounts of chlorite (prochlorite). Mineralogical analysis of the light mineral separate (specific gravity <2.7 g/cc) of the fine and very fine sand-sized fractions indicated some variation occurred in the content of K-feldspar. In general, the content of K-feldspar in the tills of their study ranged from 6 to 8% while soda-calcic feldspars constituted 8.5 to 11.5% in till from north-central parts of the province and 11.5 to 17% in till from south and south-central Alberta.

The clay mineralogy of Continental and Cordilleran till deposits in Alberta were studied in four localities in west-central and central Alberta (Twardy, 1969). His semi-quantitative mineralogical estimates indicated a predominance of montmorillonite species and illite. These two minerals alone accounted for 70 to 80% of the clay fraction. Minor amounts of chlorite, kaolinite and quartz were reported to be present. Continental till contained the highest amount of montmorillonite; the highest content of illite was present in the clay fraction of the Cordilleran till. Interstratified mica-montmorillonite was present in both tills with highest quantities associated with the Cordilleran till. The abundance of light minerals (specific gravity <2.96 g/cc) in the fine sand fraction was in most cases greater than 99% for all tills investigated. This light mineral separate was dominated by quartz with contents of soda-calcic feldspars and K-feldspar ranging from 14 to 20% and 7 to 10%, respectively. The heavy mineral assemblage of the fine sand fraction was reported to be dominated by iron oxides and opaque minerals. Amphiboles and garnet were second in dominance. Less than 5% of the heavy mineral fraction consisted of epidote, pyroxene, chlorite, zircon, apatite and tourmaline.

The nature of mica in different size fractions from three soil parent materials in Saskatchewan were described by Somasiri and Huang (1974). Their results suggested that variations in the proportion of dioctahedral to trioctahedral mica could be attributed to both a stability-size relationship and origin of the parent material. In most cases, dioctahedral and trioctahedral micas were present in the silt, sand and coarse clay fractions, while highly stable dioctahedral mica was only present in the fine clay. Furthermore, the glacial till deposits in the northern regions contained higher amounts of trioctahedral mica because of the influence of the igneous and metamorphic rocks of the Shield. In contrast, the tills of the southern regions were largely derived from local bedrock shales and contained mostly dioctahedral micas.

The discussion to this point has dealt with the mineralogical properties of bedrock and surficial materials in Western Canada. No large qualitative differences appear to exist among the different sedimentary bedrock formations of the Interior Plains. Surficial glacial deposits throughout the plains region also display qualitative similarity in terms of their mineral components and largely reflect composition of underlying local bedrock (Bayrock, 1962; Gravenor and Bayrock, 1965; Pawluk and Bayrock, 1969). In the following portion of this section, the mineral attributes of soils of the Interior Plains region will be presented and as will become apparent, soil mineral attributes will largely reflect the mineralogy of surficial and bedrock materials.

A relatively large quantity of information on the clay mineralogy of western Canadian soils was accrued and published in field tour guidebooks prepared for the Eleventh Congress of the International Society of Soil Science held in Edmonton in June, 1978 (Acton and Crosson, 1978). Review of the mineralogical information indicated mica and smectite dominate the clay fraction of well-drained soils in southern Manitoba and southern Saskatchewan. Content of smectite generally exceeded content of mica in C horizons; however, mica was more abundant than smectite in A and B horizons. In poorly drained soils from the same region, amorphous materials along with mica were dominant in clay fractions from soils developed on glacial till derived from Cretaceous shale, Paleozoic dolomitic limestone and Precambrian granites. The reported high content of amorphous material appears unusual considering bedrock and surficial deposit mineralogy. Smectite was reported to dominate the clay assemblage of soils from Alberta. Only trace amounts of mica were reported to occur in clay fractions from Alberta soils. Kaolinite and chlorite were reported to be absent or present only in trace amounts in all the pedons analyzed except for those from southern Saskatchewan which contained minor amounts of kaolinite. Vermiculite was reported to be present in trace to minor amounts in clay separates from surface horizons of Eluviated Black Chernozemic and Gleysolic soils located in the central and east-central parts of Alberta. Trace amounts of vermiculite were reported to be present in most other soils from the Interior Plains. Clay-sized quartz was described to occur in clay fractions in amazingly high amounts . For many soils, estimates of content of quartz were as high as 20 to 40% in the clay sized fraction.

The clay mineralogy of Chernozemic soils in central and southern regions of Alberta was described by Dudas (1968).

The clay mineralogical suite of A, B and C horizons was dominated by montmorillonite which constituted up to 77% of the total clay fraction. Illite (hydrous mica) was the second dominant clay mineral with percentages generally less than 25%. Chlorite and kaolinite were present in substantially lower amounts (generally less than 10%), while quartz was present in trace quantities. For some sites in the south-central region, content of montmorillonite was low with a concomitant increase in chlorite content. Some evidence was found for the presence of randomly interstratified illite and montmorillonite. St. Arnaud and Mortland (1963) examined Chernozemic and Luvisolic soils in Saskatchewan and found clay mineral suites similar to those described by Dudas and Pawluk (1970) for Black Chernozemic soils in Alberta.

The mineralogy of five geographically associated Solonetzic and Chernozemic soils developed on glacial Lake Edmonton was studied by Arshad and Pawluk (1966). Again, montmorillonite dominated the clay fraction with higher amounts of this mineral in the Bt horizons as compared to both the Ae and C horizons. Illite and kaolinite were next in abundance and chlorite and interstratified clay minerals were present in very small amounts. The type of smectite present was identified as being transitional between beidellite and nontronite. Their results for the mineralogical composition of the clay fraction were generally comparable to those reported by Dudas and Pawluk

(1970). The light mineral fraction (specific gravity <2.7 g/cc) constituted more than 95% of the 100 to 150 micrometer sand fraction and consisted of more than 70% quartz with less than 20% soda-calcic feldspars and less than 10% K-feldspars. The heavy mineral fraction of the fine sand separate was similar to that reported by Twardy (1969) for continental till deposits in central Alberta except that a substantially lower amount of hornblende was reported for the soils developed on lacustrine materials.

The mineralogy of several soils developed on glaciolacustrine materials in Alberta and Saskatchewan were studied by Rice et al. (1959). The soils included Brown and Dark Brown Chernozems from the southern portion of Alberta and Saskatchewan, Thin Black and Black Chernozems from central Saskatchewan as well as a Gray Luvisol from the Peace River district of Alberta. The coarse (2-0.2 micrometer) clay fraction was dominated by montmorillonite with the exception of clay separated from the Thin Black and Gray Luvisol soils where illite was present in higher amounts than montmorillonite. Kaolinite was present in all the profiles studied with highest amounts in the Black soil. Trace amounts of chlorite and quartz were also detected; feldspars were present in barely detectable quantities. The medium-sized (0.2-0.1 micrometer) clay fraction contained higher amounts of montmorillonite and lower amounts of illite and kaolinite as compared to their respective amounts in the coarse clay fraction. Montmorillonite again dominated the fine (<0.1 micrometer) clay fraction with only trace amounts of illite being present. The Gray Luvisolic soil was an exception as the amounts of mica approximately equaled that of montmorillonite in the fine and medium clay fractions. Chlorite was absent in both the medium and fine clay fraction from the Luvisolic soil. The fine clay from the surface horizons of the Dark Brown and thin Black soils were mostly amorphous or very poorly crystalline materials. Mineral analysis of the silt fractions indicated that quartz was the most abundant mineral component. It decreased slightly in the fine silt fraction where kaolinite and chlorite were found in appreciable amounts.

Randomly interstratified mica-montmorillonite was reported to predominate the clay fraction from seven subsoils of the prairie provinces and from soils formed on residual weathered sandstone and shales in the Porcupine Hills in Alberta (Kodama and Byrdon, 1965; Pawluk et al., 1968). It was suggested by Kodama and Brydon (1965) that many of the clays originally described as montmorillonites in soils from the Interior Plains are in fact interstratified minerals inherited from the local bedrock.

The nature of K-feldspars in soils derived from glacial till and lacustrine materials were investigated by Somasiri and Huang (1973). Potassium feldspars were reported to consist of both orthoclase and microcline with the later decreasing in quantity with decreasing particle size. Their results also suggested the presence of higher proportions of orthoclase to microcline in soils developed on lacustrine deposits as compared to those developed on till. The relative abundance of orthoclase and microcline was thought to be a reflection of the nature of the soil parent material. Earlier, Huang and Lee (1969) described the clay mineralogy of two of the above mentioned soils; a well-drained Chernozemic soil and an imperfectly drained Gleysolic soil. Their results indicated that mica and quartz dominated the coarse clay fraction (2.0-02 micrometer) with appreciable amounts of chlorite, montmorillonite, kaolinite, K-feldspar, Na-Ca feldspar and amorphous materials. The fine clay (<0.2 micrometer) was dominated by montmorillonite and mica associated with chlorite, vermiculite, amorphous materials and kaolinite.

The mineralogy of Luvisols and associated soils of the Interior plains region have been examined by many researchers. Pettapiece and Zwarich(1970) and Beke and Zwarich (1971) studying Luvisolic and Chernozemic soils from southern Manitoba found the coarse clay fraction of these soils to be dominated by illite while mixed layered illite-smectite was second in abundance. The fine clay fractions were dominated by random 40:60 mixtures of illite and smectite. Trace amounts of kaolinite and vermiculite were also reported to be present in the coarse clay fraction while those present in the fine clay fraction consisted of trace amounts of smectite, vermiculite and other mixed layer clays. However, earlier studies of similar soils developed on the Mankato till of Manitoba suggested clay mineral suites were dominated by montmorillonite and illite (Ehrlich and Rice, 1956). Other minerals present in small quantities included kaolinite, vermiculite, feldspars and quartz. The dominance of montmorillonite in the fine clay fraction of the major soil horizons was reported for five Gray Luvisols developed on till from central and north-central Alberta (Pawluk, 1961). Only small quantities of illite were reported in the fine clay fraction. On the other hand, the coarse clay fraction was dominated by illite with montmorillonite entirely absent in most Ae horizons and present only in minor amounts in B and C horizons. Authigenic chlorite was reported to occur in the coarse clay fraction from Ae horizons. Quartz was present in both the coarse and fine clay fractions in minor and trace amounts, respectively. The light mineral separate (<2.70 g/cc) of the 100 to 150 micrometer size fraction was reported to be dominated by quartz (60 to 95%) with varying amounts of soda-calcic and potassium feldspars. The content of soda-calcic feldspar (12 to 32%) was reported to be higher than the K-feldspar content (5 to 11%) with the exception of one profile where their quantities were approximately the same. The heavy mineral assembladge was dominated by iron oxides in which magnetite appeared to be altering to hematite in some B horizons. Amphiboles and garnet were second in abundance while pyroxenes, apatite, zircon, epidote, tourmaline and chlorite were present in small amounts. The clay mineralogy of the above mentioned soils appears to be similar to other soils in the Peace River area described by Lavkulich, et al. (1964). Innes (1971) studied the clay mineralogy of Luvisolic soils developed on Cordilleran till of the Foothills and on Continental till of west-central Alberta. His results suggested a qualitative and quantitative difference in clay mineral distribution occurred with profile depth. Mixed layered structures consisting of illite and montmorillonite dominated the clay fraction from Ae horizons which agrees with the findings of Pawluk (1961) who described the occurrence of similar interstratified clay assemblages in Luvisolic soils from central and northcentral Alberta. Illite and kaolinite with some amorphous materials were also present in minor quantities. The clay fraction of Bt horizons were dominated by montmorillite with significant amounts of illite and kaolinite. Preferential translocation of montmorillonite from Ae to Bt horizons appeared to account for mineralogical differences within profiles. Preferential translocation was also reported to occur in Luvisolic and other soils where lessivage has been active (Clark and Green, 1964; Arshad and Pawluk, 1966; Pettapiece and Pawluk, 1972; Brunelle and Pawluk, 1976).

Luvisolic soils developed on acidic shale in the Peace River region of northern Alberta were examined by Pawluk and Dudas (1978). Their reported clay mineralogy was quite different than the mineralogy of Luvisols described in earlier studies (Pawluk, 1961; Lavkulich et al., 1964; Innes, 1971; McKeague et al., 1972). The major clay minerals described to occur in the acid shale soils included soil vermiculte, dioctahedral mica and kaolinite. A progressive decrease in the amount of soil vermiculite associated with a concomitant increase in mica with depth was observed. Chloritic intergrades were also reported to occur in the clay-sized fraction.

The principal clay minerals present in several sandy textured Brunisolic soils from the Fort McMurray region were illite, mixed layered montmorillonite-illite, discrete montmorillonite and kaolinite in both the A and C horizons (Pawluk, 1960a). However, the clay mineral components of B horizons consisted primarily of a chlorite-like mineral together with lesser amounts of kaolinite. Other Brunisolic soils of northern Alberta developed on calcareous sandy parent material were examined by Pawluk and Lindsay (1964). Clay mineral assemblages were observed to remain uniform within profiles but displayed some qualitative differences among profiles. The coarse clay fraction from one soil consisted largely of illite, iron rich chlorite and interstratified illite-chlorite-montmorillonite with small amounts of kaolinite and quartz. The mineralogical assemblage of the other two soils consisted of illite, montmorillonite, mixed layered clay, kaolinite, chlorite and quartz. Illite and montmorillonite were dominant in the fine clay fraction of one soil while the other two soils contained approximately equal proportions of montmorillonite, illite and mixed layered clays. The rest of the minerals were present in small amounts. The principal minerals in the different sand and silt fractions included quartz, K-feldspars and soda-calcic feldspars with plagioclase twice as abundant as orthoclase(K-feldspars).

The preceding section may indicate that real and substantial differences exist in the mineralogical nature of soils of the Interior Plains. It would be logical to expect only minor variation in the clay mineral assemblages of these soils since they are all young and relatively unweathered and their parent materials are mainly derived from similar Cretaceous sedimentary rocks. Some, if not a substantial amount of the apparent variation in mineral attributes reported for soils of the plains region may be related to inaccuracies associated with qualitative and quantitative mineral analyses particularly of the clay-sized fraction. The following comparison of results obtained by different investigators for an Orthic Gray Luvisol from the Cooking Lake moraine serves to demonstrate problems associated with soil mineral analyses.

Investigators who studied the Cooking Lake soil (Pawluk, 1961; Kodama and Brydon, 1965; Twardy, 1969; McKeague et al., 1972) as well as descriptions in the ISSS tour guidebooks (Acton and Crosson, 1978) based their quantitative and qualitative clay mineral assesment on one or more of the following analyses: cation exchange capacity, X-ray diffraction, total chemical analysis, differential thermal analysis, surface area and infrared spectroscopy. In all of the investigations, X-ray diffraction analysis was used and usually at least one other method of analysis was also used to evaluate kinds and amounts of clay minerals.

The only apparent agreement among most of the investigators was on the content of montmorillonite and illite, the two most dominant types of phyllosilicates reported to occur in the Cooking Lake soil. The only exception was the quantitative estimates included in the ISSS guidebooks; montmorillonite was reported to occur in trace amounts in A and C horizons and illite was described as a trace constituent throughout the profile. Pawluk (1961) indicated content of kaolinite in the clay fraction amounted to 10 to 35% with uniform distribution of the mineral throughout the profile. Twardy (1969) estimated the content of kaolinite to be less than 10% in the clay fraction from the C horizon while McKeague et al. (1972), for the same size fraction, suggested values of 5 to 20% for C horizons and up to 40% in Ae and Bt horizons. Only trace amounts of kaolinite were reported to occur in the Cooking Lake soil in the ISSS guidebooks. Most investigators thought the amount of chlorite was small except for Pawluk (1961) who mentioned that appreciable amounts of chlorite or chlorite-like minerals were present in coarse clay separates from Ae and

Bt horizons. The presence of interstratified minerals was not mentioned by McKeague et al. (1972) nor in the ISSS guidebooks whereas Twardy (1969) found small quantities of interstratified mica-montmorillonite in the C horizon. Pawluk (1961) indicated small amounts of interstratified illite-chlorite-montmorillonite occured in the C horizon with much greater quantities of the mixed layer assemblage in the upper solum. In contrast, Kodama and Brydon (1965) found the medium-sized clay fraction consisted of 93% randomly interstratified montmorillonite-mica. Content of quartz estimated to be present in clay separates differed by as much as 40% among the various investigators. Except for the description by Kodama and Brydon (1965), vermiculite was not reported to be part of the clay mineral assemblage.

Such widely differing opinions presented by the various investigators on the nature of soil clays for a soil from the Cooking Lake area suggests a need for more accurate and standardized methods of analyses. The example just cited is probably not a unique one. In a detailed study of two Saskatchewan soils, Huang and Lee (1969) found their interpretation of mineralogical analyses did not agree with the prevailing contention that the predominant clay minerals of Canadian prairie soils are montmorillonite and illite.

It is this writers opinion that differences in qualitative and quantitative clay mineral determinations results when investigators rely heavily on X-ray diffraction analysis particularly when only a limited number of pretreatments are used prior to and during X-ray analysis. Without careful control of humidity (hydration), solvation and cation saturation, it becomes impossible to detect the presence or absence of vermiculite and randomly interstratified clays can not be differentiated from chlorite intergrades or even from smectites. Additionally, method of specimen preparation (suspension method versus paste method) for X-ray analysis can have a tremendous effect on diffraction peak intensities and the subsequent interpretations. The suspension method may cause up to 250% error resulting from coarse grained micas and kaolinite being covered by fine smectite particles causing underestimation of both mica and kaolinite (Gibbs, 1965). Lastly, several independent but standard methods of clay mineral analyses must be utilized to accurately determine the kinds and amounts of phyllosilicates present in soil.

B. Mineral Weathering in Canadian Soils

Breakdown of primary minerals in soil is largely attributable to their inherent instability in the soil environment. Mineral decomposition via chemical and biological processes in soil is accompanied by secondary mineral formation together with some loss of soluble inorganic ions. Since mineral weathering (both breakdown and authigenic formation) is a relatively slow process in the surface portions of the terrestial land mass, most soils, particularly those in temperate and cooler regions of the world, contain both inherited minerals and minerals of authigenic origin formed during the course of soil development. In many pedogenic studies emphasis is placed on the quantitative and qualitative evaluation of mineral degradation and authigenic mineral formation. To accomplish such assessments, pedologists must accurately determine the mineralogical composition of soils and segregate mineral attributes inherited from the parent material from those acquired as a result of mineral weathering accompanying soil development. To correctly differentiate inherited from acquired characteristics an investigator must determine the uniformity of the parent material and whether or not lithological discontinuities occur within the soil profile. If a soils' parent material is not uniform, it becomes exceedingly difficult to ascertain the nature and degree of mineral weathering. Before proceeding further, it should be pointed out that parent material uniformity was not evaluated in most of the mineral weathering studies to be described in the present section.

The mineral components of Canadian soils appears to be largely inherited from their parent materials (Kodama, 1979). However, some mineral weathering during pedogenesis has been reported for soils of this country. The extent of weathering in Canadian soils bears some relationship to the intensity of the soil forming processes. For this reason, pedologists (Brydon et al., 1968; McKeague and Brydon, 1970) have observed that soils belonging to different Orders in the Canadian soil taxonomic system display different degrees of weathering. The following discussion, arranged on the basis of soil orders, reviews the nature of weathering in Canadian soils.

Limited mineral weathering has taken place in most Chernozemic soils (Ehlrich and Rice, 1956; Rice et al., 1959; Dudas and Pawluk, 1970). However, mineral transformation such as illitization of montmorillonite was suggested to occur in surface horizons of both Chernozemic and Solonetzic soils (St. Arnaud and Mortland, 1963; Arshad and Pawluk, 1966; Brunelle and Pawluk, 1976). In the sand fraction, weathering found in feldspars, ferromagnesian minerals and apatite was considered insignificant (Ehlrich and Rice, 1956) or inherited from the parent material (Rice et al., 1959). Huang and Lee (1969) reported evidence for greater mineral weathering in Chernozemic soils as compared to the previous authors mentioned above. Their study suggested degradation of chlorite occurred in the Ap horizon and physical disintegration of quartz was common within the solum. Furthermore, two kinds of mineral transformations were suggested to occur in the surface horizons: (1) interlayering of hydroxy aluminum in montmorillonite of the coarse clay-fraction and (2) transformation of vermiculite to mica. The detailed mineralogical investigation of the clay fraction by Huang and Lee (1969) as compared to the less rigorous studies by the above mentioned workers may account for the apparent differences.

Mineral weathering appears to be more pronounced in Solonetzic and Eluviated Black Chernozemic soils than in Orthic Chernozems (Arshad and Pawluk, 1966; Brunelle and Pawluk, 1976). Mineral weathering reported to occur in Solonetzic and Eluviated Black soils include: (1) authigenic mica formation (illitization of montmorillonite), (2) slight weathering of K-feldspars and moderate weathering of soda-calcic feldspars, (3) degradation of montmorillonite, (4) disintegration of quartz in the Ae horizons and (5) degradation of mica in A and B horizons.

Acid leaching and consequent low base saturation, especially in the upper portions of the solum has brought about a relatively high rate and intensity of mineral weathering in Luvisolic soils. Alteration of dioctadedral micas and to a lesser extent trioctahetral chlorite has been noted in many Luvisolic soils. Alteration products include hydrous mica and smectite or vermiculite and interlayered structures and in some instances chlorite (Pawluk, 1960b, 1961; McKeague et al., 1967; Brydon et al., 1968; McKeague and Brydon, 1970; McKeague et al., 1972). Transformation of hydrous mica to montmorillonite through displacement of potassium ions was reported to occur specifically in the coarse clay fraction of Ae horizons. Mixed layered structures of illite-chlorite-montmorillonite were suggested as an intermediary component in this transformation sequence where fine clay-sized montmorillonite was the final product (Pawluk, 1960b, 1961). Luvisolic soils with low base saturation display an appreciable degree of clay mineral weathering whereas those with a relatively high base saturation are only minimally altered. For example, montmorillonite formed in the upper portion of the solum of low base saturated Luvisolics while only degradation of montmorillonite to poorly crystalline mineral material was observed in high base saturated Luvisolics (McKeague et al., 1972). Illitization of expansible minerals was also reported to occur in the surface horizons of some Luvisolics (McKeague et al., 1972).

Chlorite depletion and/or alteration to montmorillonite was suggested to occur only in a relatively intense weathering environment (McKeague et al., 1967, 1972). In most Luvisolic soils chlorite, if present, is unaltered (Brydon et al., 1968; McKeague and Brydon, 1970) or even formed in the coarse clay fraction through magnesium and/or aluminum hydroxy interlayering (Pawluk, 1960b, 1961).

Weathering of clay sized dioctahedral mica to vermiculite through depotassification and dealumination was reported to occur in Luvisolics derived from acidic shale. The strong acidic conditions was a result of oxidation of pyrite where sulphuric acid was produced. Weathering of mica was most pronounced in the upper parts of the solum of these soils as evidenced from X-ray diffraction patterns and K20 analysis (Pawluk and Dudas, 1978).

Synthesis of expanding components from amorphous materials accompanied by some degradation of primary chlorite and illite was reported to occur in Luvisolics developed on till with variable content of volcanic ash (Pettapiece and Pawluk, 1972). The expandable mineral (montmorillonite) was poorly organized and occurred with randomly interstratified materials in the Ae but displayed better crystallinity in the Bt horizons. The possible sources of the expanding clay component along with the interstratified materials were suggested to include the following: synthesis from amorphous materials, mica weathering and chlorite degradation.

Brunisolic soils developed on materials originally containing chlorite showed slight weathering of this mineral in the clay fraction to montmorillonite (Pawluk and Lindsay, 1964; McKeague and Brydon, 1970). Weathering of iron rich silt and clay-sized chlorite and slight weathering of clay-sized hydrous mica was noted by Pawluk and Lindsay (1964) for Brunisolics from northern Alberta. Transformation of biotite to vermiculite in the clay fraction was the only apparent mineral alteration in Brunisolics developed on sandy stratified till material. The clay mineralogical suite of these soils was dominated by biotite and vermiculite, most of which was inherited from nearby bedrock (Rutherford and Sullivan, 1970). Weathering of some alpine Brunisolic soils developed on materials with a large volcanic ash component was different. Intense hydrolysis of ash produced amorphous iron and aluminum materials which allegedly resulted in the formation of montmorillonite (King and Brewster, 1974). This process of montmorillonite formation would probably be comparable to that reported by Pettapiece and Pawluk (1972).

Most studies on mineral weathering in Gleysolic soils do not indicate definite weathering patterns. One study on Gleysolic soils showed mineral weathering was not detectable (Nicholson and Moore, 1978) while others indicate moderate degrees of weathering and clay mineral transformations (McKeague, 1965; Pawluk, 1971; McKeague and Brydon 1970; McKeague et al., 1971; DeKimpe 1976). Clay mineral formation or transformations suggested by the above mentioned authors include the following: (1) some chloritization of expanding lattice minerals in the Ae and B horizons (McKeague et al., 1971), (2) formation of the major proportion of expanding minerals present in the gleyed horizons (McKeague, 1965), (3) minor transformation of illite (De Kimpe, 1976) and (4) moderate weathering of chlorite in the Ae horizons (McKeague and Brydon, 1970). McKeague and Brydon (1970) pointed out that more information is required on the clay mineralogy of Gleysolic soils before any conclusions about mineral weathering in these soils can be drawn. They suggested that alternating reducing and oxidizing conditions that prevail in the surface horizons of these soils favor rapid dissolution of chlorite and that the absence of iron oxide coatings favor the weathering of mica to montmorillonite. A detailed study of Gleysolic soils revealed some weathering trends especially in Ae horizons (Huang and Lee, 1969). Mineral weathering in these soils included the following processes: (1) transformation of vermiculite to montmorillonite in the fine clay fraction, (2) degradation of both kaolinite and chlorite, (3) breakdown of montmorillonite and accumulation of amorphous alumino-silicates in both the coarse and fine clay fractions, (4) decrease in the content of potassium feldspars and concurrent increase in the content of mica in the coarse clay fraction, and (5) accumulation of quartz particularly in the Ae horizon.

The greatest degree of mineral weathering in Canadian soils is observed in soils of the Podzolic Order. Destruction of chlorite in Ae horizons of Podzols has been described by many workers (Kodama and Brydon, 1968; Brydon et al., 1968; McKeague and Brydon, 1970; De Kimpe, 1976; Nicholson and Moore, 1978). In contrast, chlorite in B horizons of Podzolics is thought to be protected from weathering reactions by iron and aluminum coatings (Brydon et al., 1968). Other mineral weathering products found in clay fractions from Podzolic B horizons include chlorite-like minerals or chloritized vermiculite (Pawluk, 1960a, 1963; McKeague, 1965; McKeague and Brydon, 1970). Chlorite formation in B horizons of Podzolics was suggested to involve interlayering of degraded illite by aluminum hydroxy materials or from migration of Al^{+++} ions into interlayer positions of montmorillonite with subsequent crystallization of gibbsite layers. Interstratified minerals were also identified in B horizons as intermediate products of mica degradation (De Kimpe, 1974).

Formation of smectite or expanding lattice minerals is often reported in Ae horizons of Podzolic soils (McKeague, 1965; Brydon et al., 1968; McKeague and Brydon, 1970; De Kimpe, 1974; Nicholson and Moore, 1978). Smectite formation was regarded by some researchers a result of mica transformation.Others suggested its synthesis from amorphous materials (Brydon et al., 1968) or even its formation at the expense of kaolinite or mica (Nicholson and Moore, 1978). Formation of kaolinite in Ae horizons of Podzols was reported in two instances (Mutwewingabo, 1975; DeKimpe, 1976). DeKimpe (1976) suggested conditions of low base saturation and extreme acidity in Ae horizons would favor the formation of kaolinite.

Podzolic soils developed on parent materials containing variable amounts of volcanic ash display some differences in the nature of weathering processes and products. Chloritization of vermiculite was suggested to occur in surface horizons of Podzolic soils containing volcanic ash (Sneddon et al., 1972; Pawluk and Brewer, 1975). Chloritization of vermiculite was a result of accumulation of amorphous materials from the readily weatherable volcanic ash. Formation of organo-aluminum complexes prevented crystallization of inorganic materials as clay minerals (Pawluk and Brewer, 1975). Mica transformation to montmorillonite and its further degradation to kaolinite was suggested to occur in Ae horizons (Sneddon et al., 1972). These authors also described formation of allophane in B horizons.

C. Geology, Climate, Topography, Drainage, Vegetation, Parent Materials and Soils of the Sand River Area.

Sites selected for this study are located in the general area referred to as the Edmonton-Lac La Biche-Cold Lake area (Figure 1). Five out of six sites are located in the Sand River map sheet (Figure 2) whereas the sixth site (site 6) is located in the Cooking Lake moraine near Edmonton (Figure 1). In the Sand River area four sites (sites 1,2,3 and 5) are located in the Cold Lake-Bonnyville region while one site (site 4) is located near Lac La Biche.

Glacial deposits of the Edmonton-Lac La Biche-Cold Lake area are underlain by the Horseshoe, Bearpaw, Belly River and Lea Park Formations of Upper Cretaceous age (Figure 1). The Horseshoe Canyon Formation underlying the Cooking Lake area consists of non-marine sandstone, shale and coal while the Bearpaw Formation consists of marine shales and minor sandstones. The Belly River Formation present in the south west corner of the Sand River map sheet which includes the

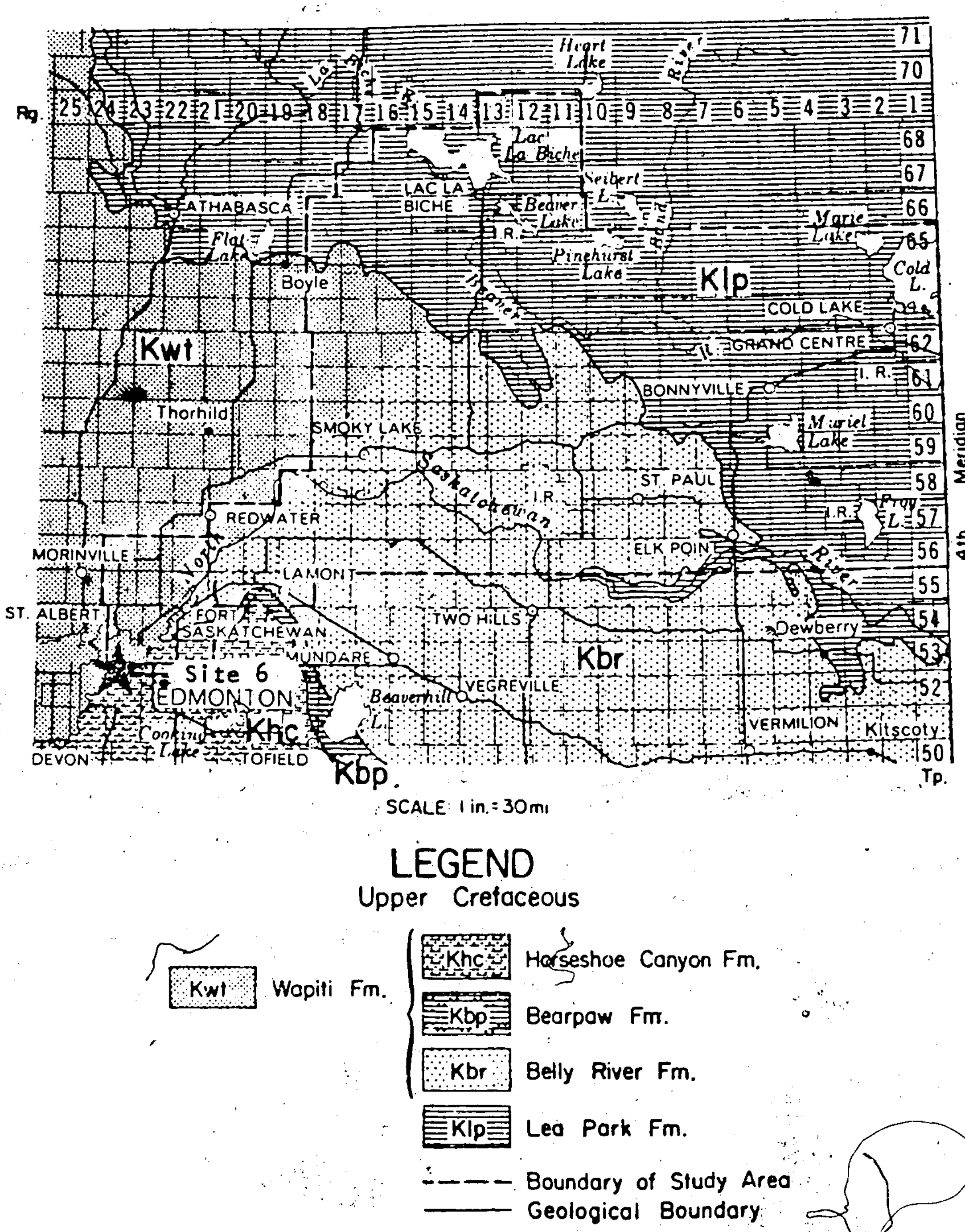


Figure 1. Bedrock geology of Edmonton – Lac La Biche – Cold Lake Area (From Yoon and Vander Pluym, 1974)

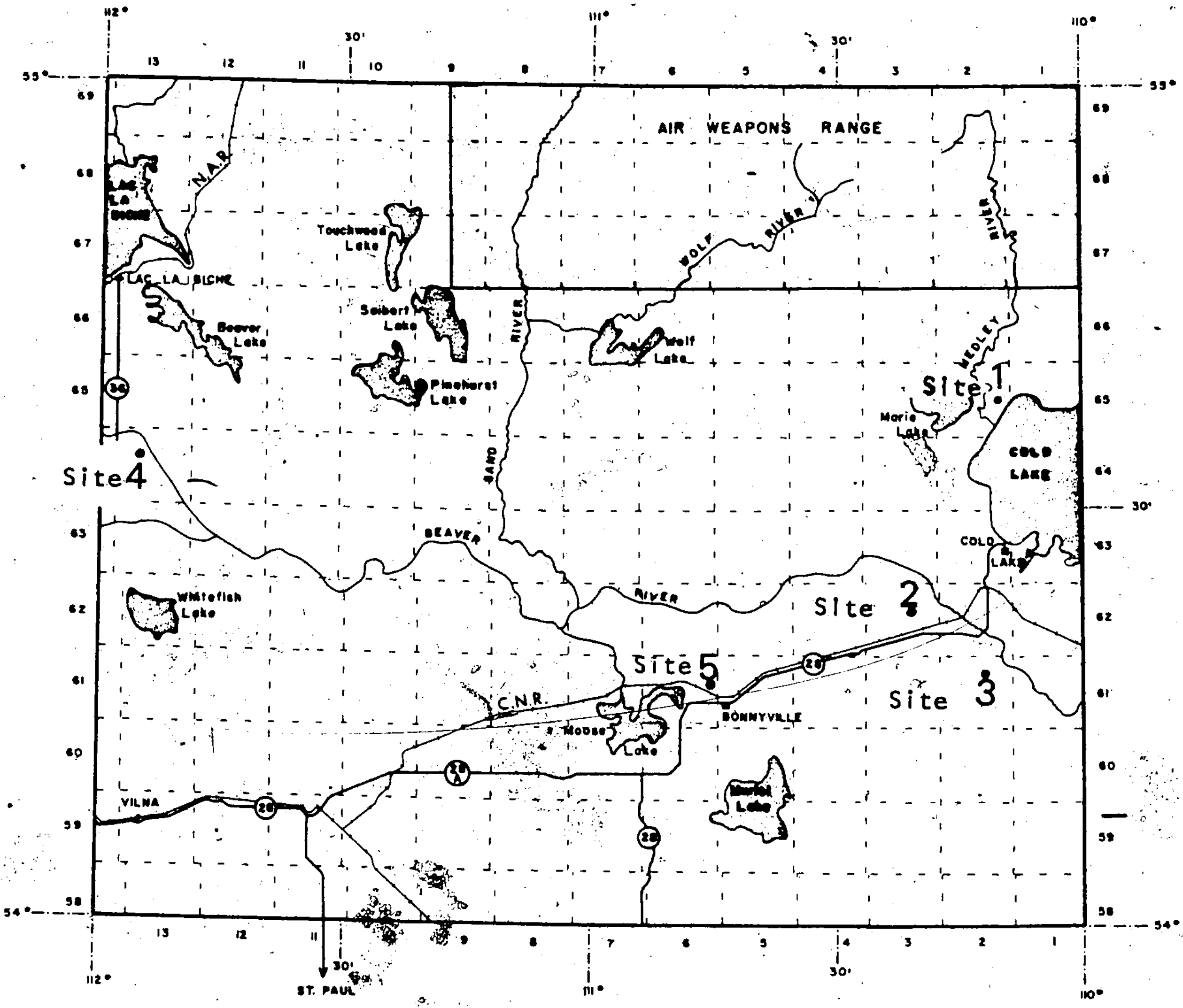


Figure 2. Map showing sampling sites in the Sand River area

Cold Lake-Bonnyville region consists of non-marine grey to greenish-grey, thick-bedded feldspathic sandstone, grey clayey siltstone, grey and green mudstone and concretionary ironstone beds. The Lea Park Formation consists of massive silty dark grey marine shale and argillaceons siltstone with some fine grained sandstone lenses.

The major ice advances into the Edmonton-Lac La Biche-Cold Lake area were from the Canadian Shield in the northeastern corner of Alberta; the rocks of the Shield consists of granite gneiss (>50%), porphyritic and massive granite and metasediments of Precambrian age. Carbonates and evaporite rocks of Devonian age adjacent to the shield rocks extend from the southeast to the northwest in the northeast corner of Alberta almost perpendicular to the major ice advance. The glacial history of the Sand River Area consisted of four advances of glacial ice from the northeast depositing four separate till units. As the first and second ice sheets retreated to the east, they were followed by short and undetermined interglacial periods. The third advance covered the eastern half of the area and deposited a distinctive high carbonate till. It was followed by an extended interglacial period during which an extensive weathered zone formed. The final ice advance extended past the western boundary of the area depositing the surficial till from which the present topographic surface evolved (Gold, 1978).

There is little agreement among various geologist on the age correlation of the glacial deposits in Alberta (Gravenor and Bayrock, 1965). It appears that the last major glacial advance was more than 31,000 years ago and retreated about 11,000 years ago. Following the ice retreat and deposition of the glacial drift, the climate reached an optimum about 5000 years ago with a slightly cooler present day climate.

The climate of the Sand River area lies within the continental climatic zone characterized by cold winters and warm summers. The mean annual temperature within the area varies from 0 to 1.5°C and the mean Jan, April, July and October temperatures vary from 0 to -16°C, 5.6 to 11.1°C, 16.7 to 22.2°C and 0 to 5.6°C, respectively. The warmest month is July with an average maximum of over 26.7°C. The coldest month is January with an average minimum of minus 23.4°C. The frost free days range from 90 to 100 days in the southern portion of the Sand River area and decreasing to about 60 to 80 days in the north. The mean annual precipitation varies from 457 to 508 mm at Cold Lake to 406 mm to the extreme west of the Sand River area. Mean annual snow fall varies from 1,016 to 1,270 mm at Cold Lake to 2,032 mm in the extreme north of the area. The climate of the region is similar to the Cooking Lake area near Edmonton although temperatures are milder at Cooking Lake with a frost free period of up to 120 days (Atlas of Alberta, 1969).

Five of the six study sites used in this investigation are located within the Sand River map sheet (73-L). The north and north central part of the map area lies within the Mostoos Hills Upland having an elevation of generally over 660 meters with gently undulating topography consisting of morainal plains and strongly rolling hummocky moraine (Atlas of Alberta, 1969). The general slope of the area is gradual from the north and southwest to the central and east-central parts of the area.

Beaver River and its tributaries (Sand, Amisk and Moose Lake Rivers) constitute the major drainage system of the area. Additionally the Owl River, North Saskatchewan River and Medley River drain the extreme three corners of the map area (Kocaoglu and Brunelle, 1975).

The Sand River area (map sheet 73-L) lies within the Boreal Forest region with the exception of its southern margin which lies within the Boreal Forest-Parkland transition. Boreal forest consists of mixed wood vegetation dominated by aspen poplar and balsam poplar. Jack pine is dominant on well drained sandy sites while tamarack and black spruce are prevalent in poorly drained areas. Understory vegetation consists of shrubs and grasses.

The main soil parent material in the Sand River area is glacial till (Figure 3). The glacial till deposits include hummocky, undulating, rolling, ridged and fluted morainal plains. The till in the western portion of the area is

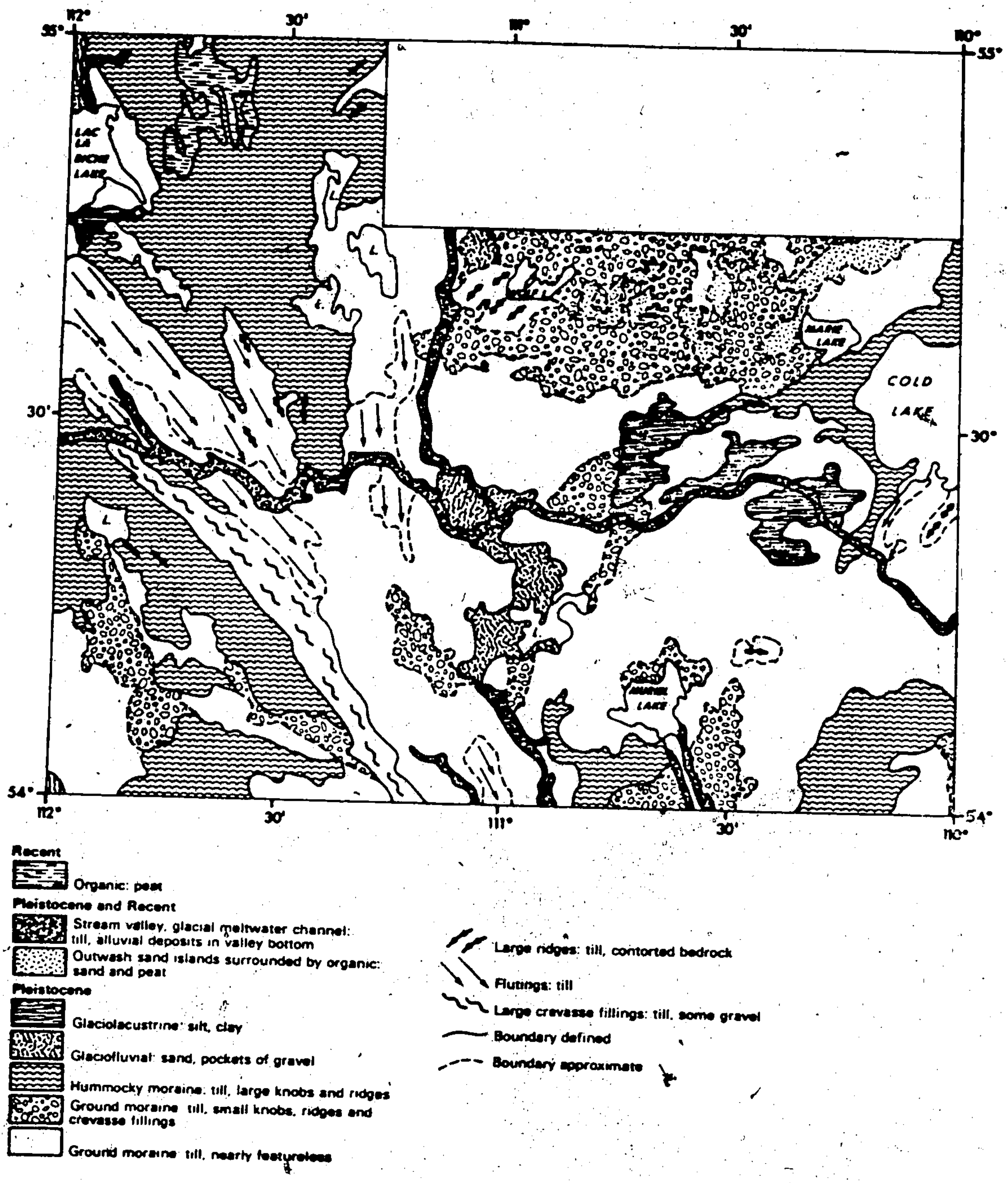


Figure 3. Surface deposits and soil parent materials of Cold Lake - Bonnyville area (from Kocaoglu and Brunelle, 1975)

composed of medium to fine textured, weakly calcareous compact till while that in the remainder of the area is of high lime content and friable consistency. Relatively small areas are covered with glaciolacustrine silts and clays. These almost flat areas were former low lands flooded by glacial melt waters. Other deposits include glaciofluvial sands and gravels and alluvial deposits adjacent to the main rivers and streams.

Luvisolic soils developed under forest vegetation comprise approximately 60 per cent of the area. Chernozemic soils dominate the discontinuous grassland-forest zone in the southern part of the area. The aerial extent of Chernozemic soils was estimated to be 9 %. Soils belonging to the Eluviated Black and Orthic Dark Gray subgroups were the dominant Chernozemic soils formed in the area. Both Degraded Chernozemic and Dark Gray Luvisolic soils probably developed under forest vegetation with the former soils being influenced to a great extent by influx of prairie vegetion. Brunisolic soils occupy approximately 9 per cent of the area. The distribution of Brunisolic soils coincides with the distribution of sandy glaciofluvial deposits. Gleysolic soils, formed in poorly-drained depressional positions constitute approximately 5 per cent of the area. Organic soils were found mainly in the northern parts and comprise approximately 12 per cent of the area surveyed (Kocaoglu and Brunelle, 1975).

III. MATERIALS AND METHODS

A. Field Sampling

Representive pedons were selected after examining several sites. The objectives in sampling were to select profiles representing soils developed on glacial till deposits in well drained positions on the landscape as well as representing the most dominant soil and landscape features present in the Cold Lake-Bonnyville area. In order to conduct a comparative study, a pedon of an Orthic Gray Luvisol was also sampled in the Cooking Lake area near Edmonton. The five profiles sampled in the Sand River area included two Orthic Gray Luvisols, two Dark Gray Luvisols and one Eluviated Black Chernozem.

Site locations are shown in Figure 2. Detailed site information and morphological descriptions are included in the appendix . Samples were collected from each horizon recognized in the field. Soil peds were collected from the main genetic horizons other than the Ae horizon for micromorphological studies. The samples were dried, gently crushed and passed through a 2 mm sieve. The fine earth fraction of each sample was thoroughly mixed and samples were stored for analyses.

B. Soil Physical and Chemical Analyses

Particle size analysis by the pipette method and pH measurements in distilled water and in 0.01 M CaCl2 were conducted according to methods outlined in Manual on Soil Sampling and Methods of Analysis (McKeague, 1978).

Total cation exchange capacity for the whole soil was determined by the NH4OAC method(McKeague, 1978); displaced cations were determined by atomic absorption spectroscopy (Pawluk, 1967).

Calcium carbonate equivalent was determined according to the method described by Bundy and Bremner (1972). Carbonate determinations were conducted for the whole soil, sand, silt and clay fractions from horizons of lime accumulation. The determination was conducted for the individual fractions to determine the distribution of lime among these fractions.

Total carbon was determined for both the whole soil and the clay fractions from the major horizons of organic matter accumulation by wet oxidation (modified Walkley-Black) as outlined by McKeague (1978).

C. Mineralogical Analyses

a. Fractionation and pretreatment of the samples

Size fractionations were conducted on 25 to 75 grams of <2 mm soil. Soils were dispersed in 250 ml of distilled water by ultrasonification for three minutes at 400 watts using a Braunsonic 1510 vibrator probe(Edwards and Bremner, 1967). Flocculation of calcareous samples was avoided by using 10 to 20 grams of soil during the sonification treatment. The dispersed soil samples were then passed through a 300 mesh sieve (0.047 mm) to separate sand fractions. Total clays were collected from the remaining suspensions using gravity sedimentation as outlined by McKeague (1978); separated clays were flocculated using N CaCl2 solution. Flocculated clays were then washed free of excess CaCl2 by centrifuge washings with distilled water after which they were freeze-dried and stored for subsequent analyses.

The total silt fraction remaining after removal of all clay was dried and split into two subsamples. Carbonate and organic matter were removed from both sets of subsamples as described by Jackson (1975) after which they were washed free of NaOAc and H2O2 by distilled water. Following removal of carbonate and organic matter, silt fractions were again sonified and all released clay was removed by gravity sedimentation to ensure silt separates contained little or no clay. One set of silt subsamples was transfered to centrifuge tubes, washed once with 95% ethanol and twice with 99% ethanol after which the samples were dried at 70°C and saved for determination of vermiculite. Fractionation of the silt into fine (2 to 5 micrometer) and coarse and medium (5 to 50 micrometer) sizes was conducted on the other set of subsamples as outlined by McKeague (1978).Both fine and coarse and medium size fractions were saturated with NH_4^+ using N NH_4OAC. The excess NH_4OAc was washed once with 95% ethanol, twice with 99% ethanol and then dried at 70°C and saved for X-ray diffraction analysis and heavy and light mineral separations.

Sands were further fractionated into the fine (0.05 to 0.25 mm) and coarse plus medium (0.25 to 2 mm) size fractions using a 60 mesh sieve. Fine sand fractions were treated for removal of organic matter and those from horizons of lime accumulation were treated for removal of carbonates. Both fine and coarse plus medium sand fractions were saved for heavy and light mineral separations.

b. Clay mineral analyses

Cation exchange capacity of freeze dried clays (<2 micrometer fraction) was determined as described by McKeague (1978). The clays were subjected to a sonification treatment for about 30 seconds during the final saturation and washing steps in order to obtain complete saturation and thorough washing.

Surface area of freeze-dried clays was measured by the ethylene glycol monoethyl ether adsorption technique as outlined by Carter et al. (1965).

The method of Coffman and Fanning (1974) for determination of vermiculite in whole soils was utilized with slight modification in this study for soil clays. Sodium saturated dry clay (0.4 to 0.6 g) previously treated for removal of carbonates and organic matter was used for the determination. A short sonification treatment was incorporated into the procedure during the last saturation and ion displacement steps.

Total dissolution of NH_4^+-saturated clays was conducted as outlined by Pawluk (1967). Following sample ignition at 850°C and dissolution using HCl-HF, potassium was determined by atomic absorption and the K_2O content was calculated to determine content of mica in the clay fraction.

Slides of Ca^{++} and K^+-saturated clays were prepared for X-ray diffraction analysis using the paste method of Theisen and Harward (1962). As had been demonstrated, the more common suspension method of slide preparation is considerably inferior to the paste method (Theisen and Harward, 1962; Gibbs, 1965). X-ray diffractograms were obtained using a Philips diffractometer equipped with a curved crystal monochrometer using CuK_α radiation generated at 40 KV and 20 MA. X-ray diffractograms of calcium-saturated slides were obtained after they were equilibrated at 54% R.H., with ethylene glycol and with glycerol as outlined by Chichester et al. (1969). For potassium-saturated slides, X-ray diffractograms were first obtained for specimens treated to 105°C. The relative humidity within the sample chamber of the diffractometer was maintained at 0% during the analysis. These same specimens were then equilibrated at 54% R.H. and again subjected to

X-ray analysis to evaluate degree of rehydration to assist in the interpretation of the presence or absence of vermiculite. Diffractograms were also obtained for K-saturated specimens heated at 300°C and at 550°C according to the methods outlined by Chichester et al. (1969).

c. Silt mineral analysis

Determination of amount of vermiculite in silt fractions was accomplished using the method of Coffman and Fanning (1974). From 1 to 3 grams of silt previously treated for removal of carbonates and organic matter were used in the determinations. Content of vermiculite was calculated on the basis of the difference between the calcium exchange capacity and the potassium exchange capacity.

For heavy and light mineral separations, 1 to 5 grams of the 5 to 50 micrometer silt fraction previously treated for removal of carbonates and organic matter was separated into the <2.59, 2.59 to 2.72 and >2.72 g/cc specific gravity separates using the centrifuge method of Jackson (1975). The recovered mineral separates were washed with acetone and dried at 70°C prior to subsequent analyses. The purpose of the heavy and light mineral separation was to concentrate in a relatively pure form K-feldspars in the <2.59 g/cc separate, quartz and plagioclases in the 2.59 to 2.72 g/cc separate and heavy minerals including mica in the >2.72 g/cc separate. A portion (0.3 to 1 g) of each of these density separates was crushed using an agate mortar and pressed into a powder slide holder having a 7x5x2 mm sample cavity. X-ray patterns were obtained using a Philips diffractometer utilizing CuK$_\alpha$ radiation and a scanning speed of 1/3 ° 2θ/min. The diffractograms were used for mineral identification and for quantitative determination of orthoclase and microcline. The latter determination was conducted using a standard curve prepared from diffractograms obtained for known mixtures of standard Madagascar orthoclase and standard microcline from Perth, Ontario.

The <2.59 and the 2.59 to 2.72 g/cc specific gravity separates of the silt fraction previously used for X-ray analysis were then used for total chemical analysis by the HCl-HF dissolution method (Pawluk, 1967). The oxide percentages of Na, K, Ca, Mg, Fe and Al were calculated following atomic absorption determinations. Values obtained for Na2O, K2O and CaO in the two density separates were used to calculate the mole per cent of albite, orthoclase and anorthite. The theoretical aluminum content of samples were then calculated on the basis of formula weights and the known quantities of albite, orthoclase and anorthite obtained from chemical analyses for Na, K and Ca. These theoretical aluminum values were then compared to the measured aluminum contents found in the HF-HCl digests to assess the accuracy of the allocation of Na, K and Ca to the respective feldspar mineral. The MgO and Fe2O3 values were used to assess the level of impurities in the feldspar and quartz separates.

Selected samples of silt separates from the major genetic horizons and specific gravity separates were examined by scanning electron microscopy. Samples were smeared on cover slips using acetone, dried and then sputtered with 150 Å gold. Electron micrographs were obtained using a Cambridge S150 scanning electron microscope. The purpose of this examination was to observe the surface morphology of feldspars and quartz in order to gain some additional information on morphological features related to weathering.

d. Sand mineral analysis

Appropiate mixtures of heavy liquids were used to fractionate both fine sands (0.05 to 0.25 mm) and coarse plus medium sands (0.25 to 2 mm) into <2.53, 2.53 to 2.59, 2.59 to 2.72 and >2.72 g/cc specific gravity separates (Jackson, 1975). Fine sand fractions were treated for removal of organic matter and carbonates prior to the heavy liquid separations; coarse and medium sand fractions received no pretreatments prior to heavy liquid separations. Fine sand density separates were then analyzed for total elemental composition and by X-ray diffraction and electron microscopy as previously described for silt separates.

Petrographic analysis was conducted for the total sand fraction and for the specific gravity separates (>2.53, 2.53-2.59 and <2.72 g/cc) of both the fine and coarse plus medium sand. Grain mounts were prepared using the method described by Innes and Pluth (1970).

IV. RESULTS AND DISCUSSIONS

Inorganic constituents in the sand, silt and clay fractions will be presented to document the mineralogical characteristics of the soils in order to meet the first and second objectives of this study. Uniformity of the parent materials will then be considered to determine whether or not lithological discontinuities occur within the soil pedons. The nature and extent of mineral weathering accompanying pedogenesis will then be presented for pedons having a uniform parent material.

Information on the chemical and physical properties of soil is essential in understanding prevailing pedogenic processes. The important pedogenic processes in Luvisolic and Chernozemic soils include humification, translocation by solution, colloidal clay translocation and various associated physical, chemical and biological weathering processes. The first three processes will be discussed in association with the presentation of the physical and chemical properties of the soils. In accordance with the main objectives of this study and the sequence previously discussed for presenting the data, the results and discussion section will be divided into four parts: (A) physical and chemical soil properties and their pedogenic implications, (B) mineralogy of the sand, silt and clay fractions, (C) parent material uniformity and (D) mineral alteration during pedogensis.

A. Physical and Chemical Soil Properties and Their Pedogenic Implications

a. Particle size distribution

Particle size distribution in the soils of this study is affected to a great extent by clay translocation. Migration of clay in youthful soils is more likely to occur subsequent to leaching of soluble salts and carbonates afterwhich wetting will facilitate eluviation and drying of the resultant suspension will cause illuviation of suspended material at depth (McKeague and St. Arnaud, 1969). Lessivage or clay migration without undergoing chemical transformation has been considered esssential for the formation of argillic horizons (Duchaufour, 1970). Various chemical and physical aspects of the mechanism of clay migration are beyond the scope of this study but could be found elsewhere (Bloomfield, 1954; DijKerman et al., 1967; Smeck et al., 1968; Allan and Hole, 1968; Oertel, 1970; Miller et al., 1971b; Ritchie et al., 1974; Mel'Nikova and Kovenya, 1974a and b). Understanding the mineralogical implications of this important pedogenic process requires an analysis of it's extent in the soils under investigation. This analysis is partly achieved by examination of particle size distribution data, morphology in the field and micromorphological evidence in thin sections.

The texture of the parent material of the profiles is clay loam with the exception of sites 5 and 6 which have loam to sandy clay loam textures (Tables 1 to 6, Appendix A). Based on particle size data, the tills of all profiles studied contain 24 to 27% silt with a generally similar coarse, medium and fine silt distribution. A recognizable difference occurs in the sand and clay content. Their distribution suggest till could be separated into four groups: site 4, site 1, sites 2 and 3, and sites 5 and 6. The percentage of sand in till for these four groups increases from 34 to 51% while the per cent clay decreases from 39 to 24% and the percentage of fine clay ranges from 22 to 16%. The till textural groups are readily recognized in the cumulative-frequency curves (Figure 4). However, if particles finer than fine sand are considered, the curves indicate only three textural groups (4, 1 and the remaining four sites) since tills from sites 2, 3, 5 and 6 contain relatively similar amounts of silt and clay-sized particles. The relatively high sand content at site 4 and site 6 as well as the differences in the content of coarse, medium and fine sand separate them into four possible textural groups.

Variation in texture of glacial till is a function of local bedrock, direction of ice movement, nature of preglacial deposits and mode of deposition (Gravenor and Bayrock, 1965). The recognized landforms of the sampling sites were mainly formed subglacially. Hummocky moraine landform is believed to be isolated deposits resulting from slow disintegration of stagnant ice that contained considerable till in its lower layer while morainal plain

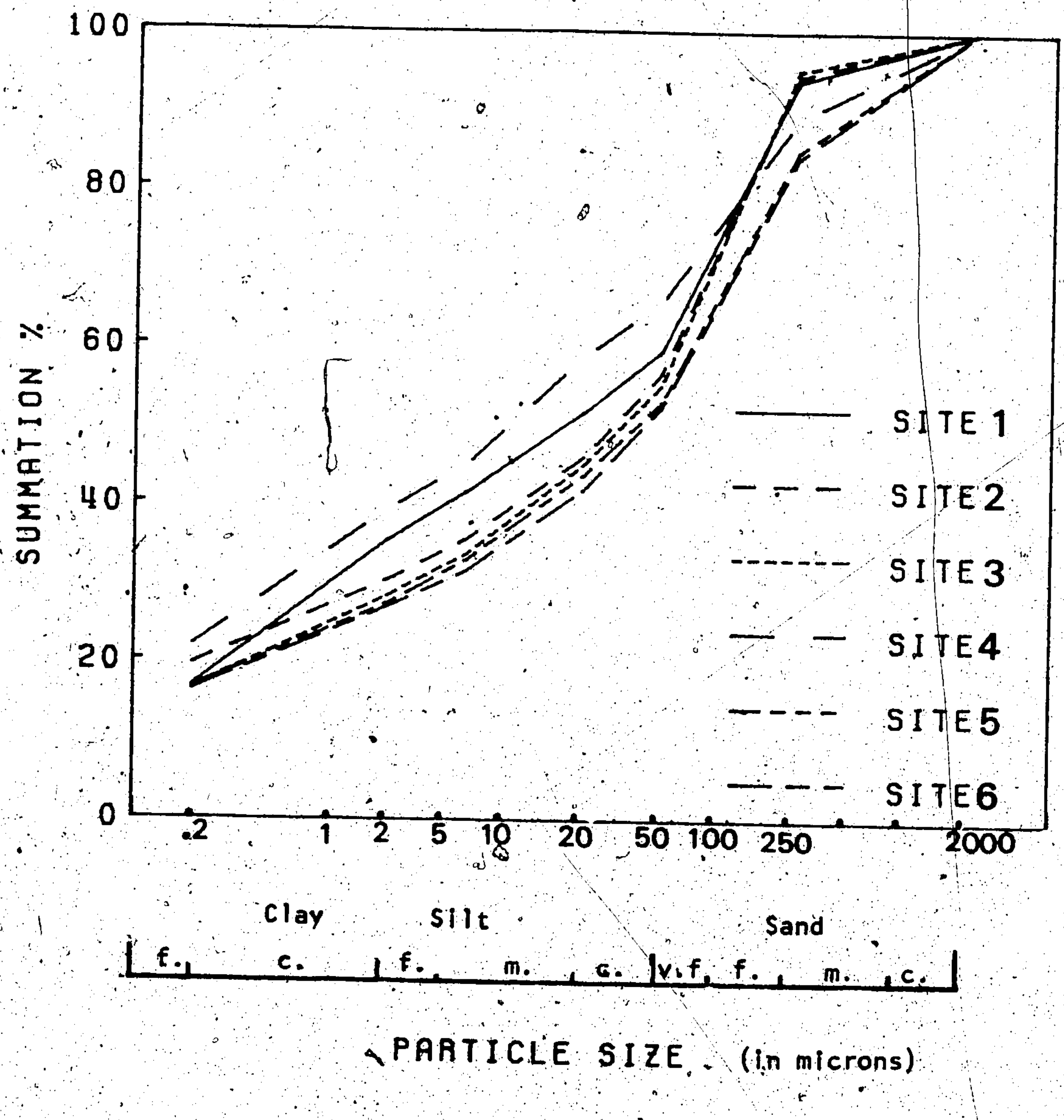


Figure 4. Particle size cumulative-frequency curves representing the parent material of the six study sites.

deposits are comprised of ridges and depressions of low relief (Bird, 1972). The hummocky moraine is believed to contain high amounts of sand or gravel and may include stratified tills (Clayton et al., 1977). Sites 2, 3 and 5 were on undulating or rolling morainal plains while site 4 was on a ridged morainal plain and sites 1 and 6 were on hummocky moraine (site descriptions, Appendix A). The high content of clay at sites 4 and 1 would probably be due to incorporation of large amounts of fine textured material from shale bedrock while the high percentage of sand in the Cooking Lake till (site 6) may be related to mode of deposition. Similar textural results for the Cooking Lake till was reported by Twardy (1969) but he related the relatively high sand content to both mode of deposition and presence of sand pockets. Sand pockets were not evident in the site sampled for this study.

Variations in the content of sand within profiles (Figures 5 to 7) are mainly a reflection of clay translocation. A two fold increase in the content of sand in the Ae horizon (relative to the C horizon) of site 1 is probably a reflection of both the eluviation process and the presence of sand pockets which were evident in this profile at the time of sampling. The hummocky morainal plain land unit described in the soil survey report of the Sand River area (Kocaoglu and Brunelle, 1975) was mentioned to contain sand and gravel pockets as well as layers of sand capping the till deposits.

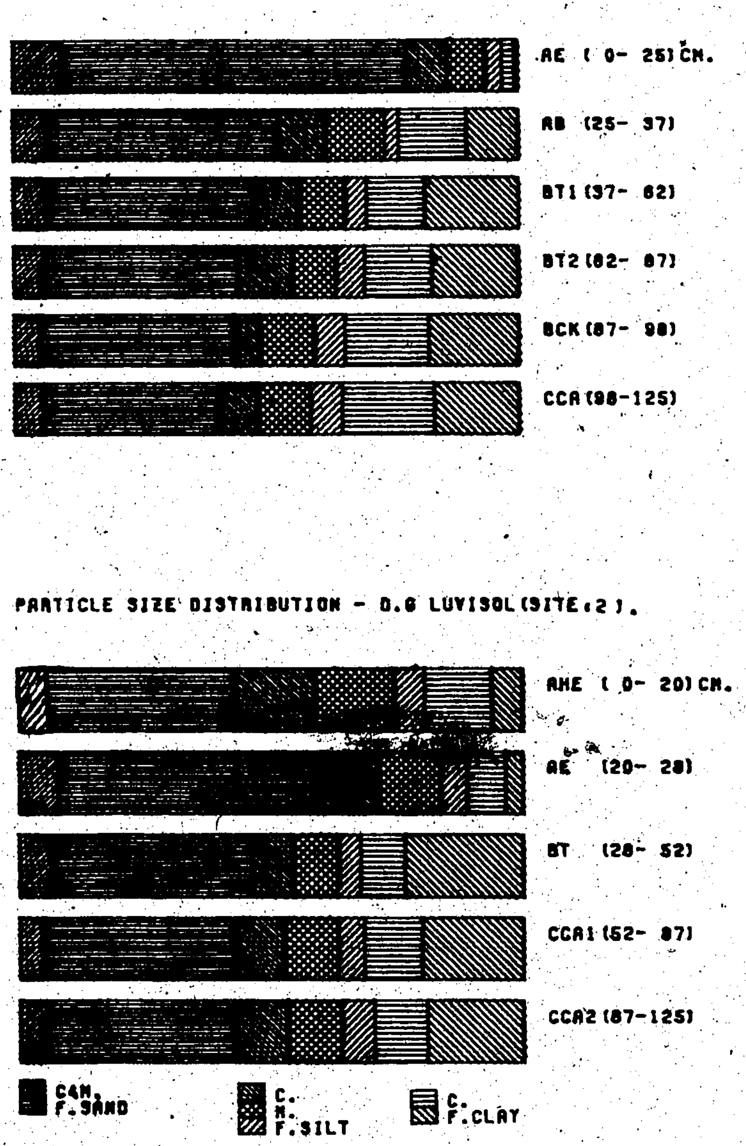


Figure 5. Particle size distribution for an Orthic Gray Luvisol (site 1) and a Dark Gray Luvisol (site 2)..............................

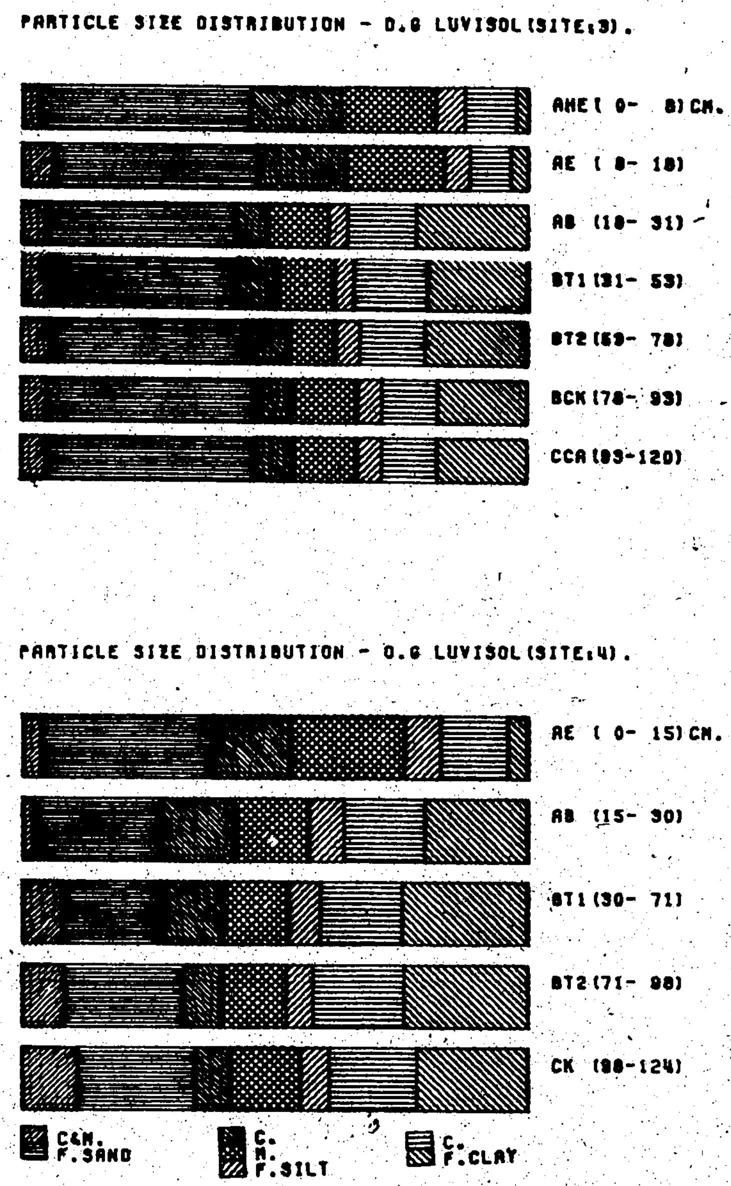
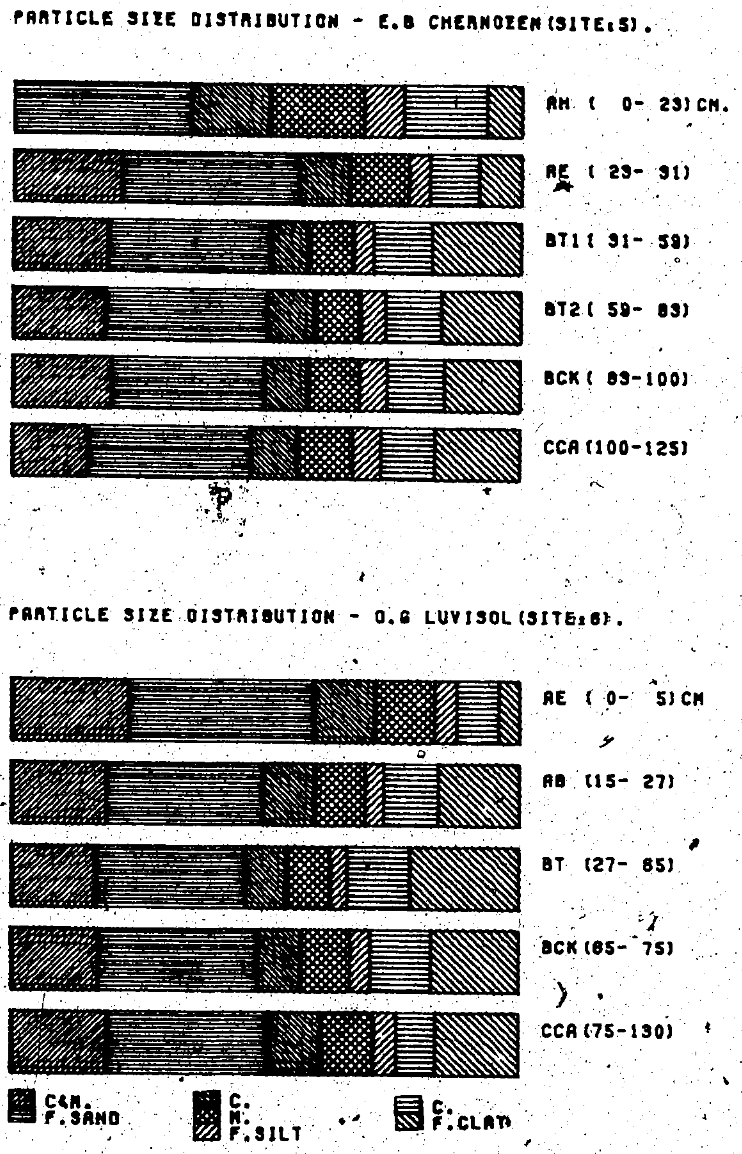


Figure 6. Particle size distribution for a Dark Gray Luvisol (site 3) and an Orthic Gray Luvisol (site 4)..


PARTICLE SIZE DISTRIBUTION - E.B CHERNOZEM (SITE:5).

PARTICLE SIZE DISTRIBUTION - O.G LUVISOL (SITE:6).

Figure 7. Particle size distribution for the Eluviated Black Chernozem (site 5) and an Orthic Gray Luvisol (site 6)..............................

Silt distribution within profiles shown in Figures 5 to 7 are largely a function of the eluviation of clay. Ratios of silt to clay contents (Tables 1 to 6, Appendix A) also display the eluviation process. Ratios range from 1.5 to 5.7 for eluviated horizons and are generally less than 1 for Bt and C horizons. The distribution of fine silt is relatively uniform with depth for each profile (Figures 5 to 7). However, a 3% increase in fine silt occurs in the Bt horizon of profile 1 and a 1% increase in the Bt horizon of profile 3 in comparison to respective contents in AB horizons and, if significant, may reflect minimal translocation of fine silt-sized particles in these profiles (Wright and Foss, 1968). The distribution of medium and coarse silt in profiles 2,3,4 and 5 indicate a 5 to 15% enrichment in Ah, Ahe and Ae horizons. This may be a result of clay translocation from these horizons (negative enrichment). To assess this possibility, particle size distribution was recalculated on a clay-free basis. The enrichment previously noted maintained at a relatively high value (about 10%) when the non-clay fraction of the surface horizons was considered and suggests the possible role of other process such as the physical breakdown of sand-sized particles to silt size and/or eolian additions to the soil surface. The increase in medium and coarse silt particles in the surface horizons is further illustrated in representive cumulative curves (Figures 8 and 9). Similar figures can be found in the appendix for the remaining soils (Figures 1 to 4, Appendix

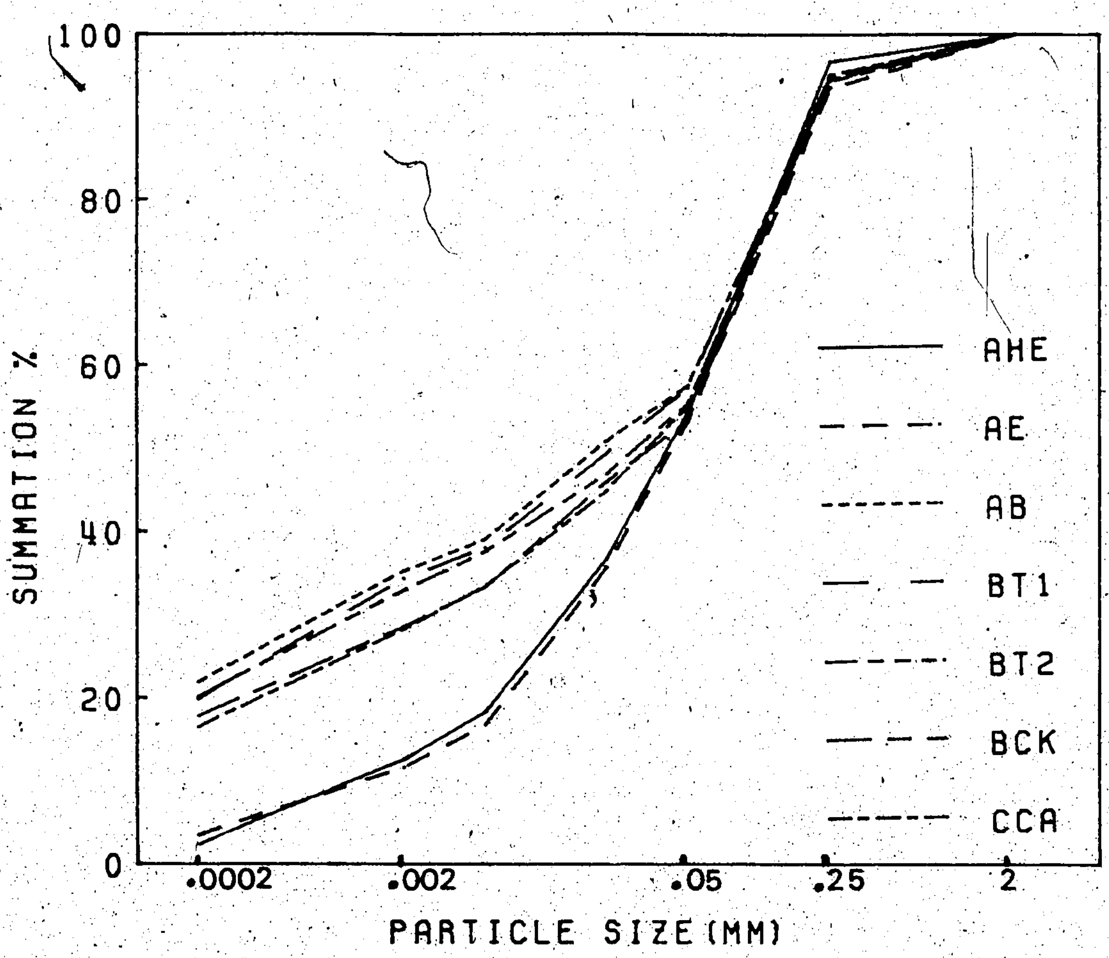


Figure 8. Particle size cumulative-frequency curves representing horizons at site 3..............

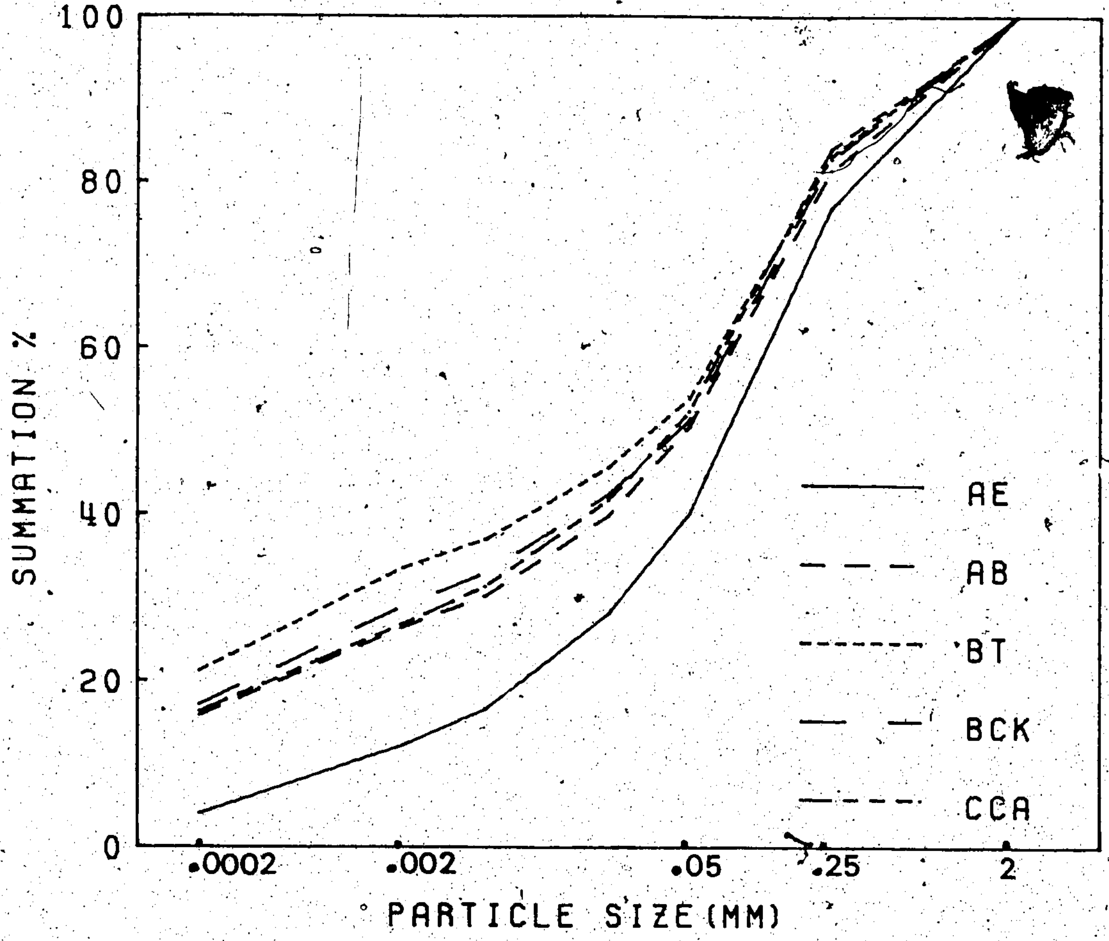


Figure 9. Particle size cumulative-frequency curves representing horizons at site 6..............

A). The curves show Ae, Ahe and Ah horizons contain less clay and/or more silt sized particles as compared to other horizons. As was previously mentioned, there was little or no evidence to indicate translocation of fine silt in soils from sites 2,4,5 and 6. In general, it appears that the distribution of silt is related to negative enrichment through clay eluviation, physical disintegration of coarse particles and eolian additions.

Distribution of clay within the profiles (Figures 5 to 7) shows an abrupt increase of 9 to 27% from A to B horizons. This increase yields a Bt/Ae clay ratio range of 1.6 to 9 (Table 1 to 6, Appendix A) and reflects the eluviation process. The range in content of clay in Bt horizons is 27 to 43% compared to a range of 24 to 35% in C horizons. This yields Bt/C clay ratios of 0.9 to 1.2. All profiles, except the one from site 1, displayed ratios greater than unity and indicates higher contents of clay in B horizons as compared to C horizons. The narrow range of Bt/C clay ratios (0.9 to 1.2) suggests eluviated clay may have been leached beyond the B horizons. Fine clay is somewhat enriched in the upper Bt horizons of all profiles compared to C horizons (Figures 10 to 15 and Tables 1 to 6, Appendix A). Since total clay distribution within a profile may also be affected by uniformity of the parent material, it follows that fine clay to coarse clay ratios and/or the overall calculated quantities of gains or loss in fine clay would probably be the best analytical indicators of clay

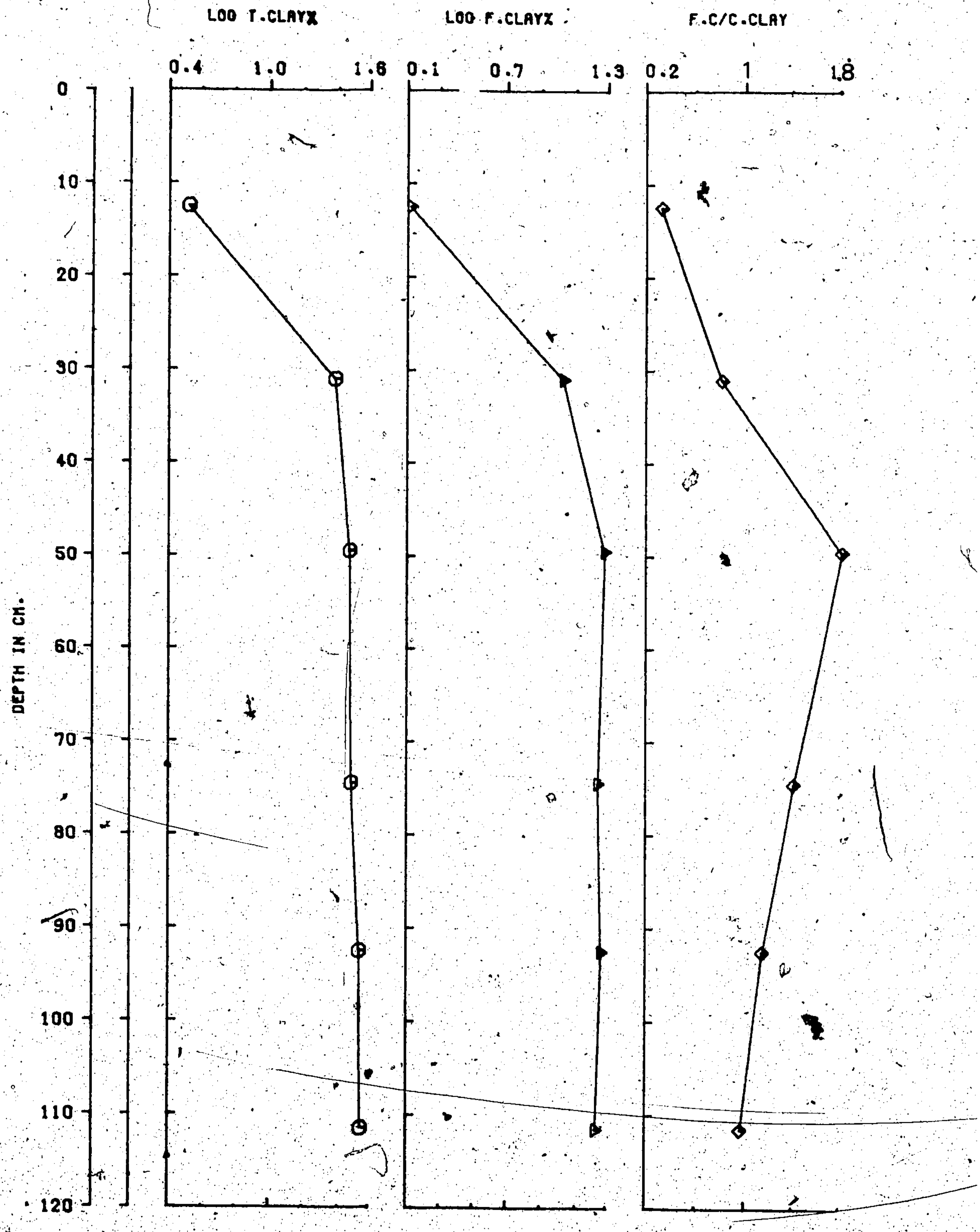


Figure 10. Total clay and fine clay content and fine clay:coarse clay ratio with depth for an Orthic Gray Luvisol (site 1)

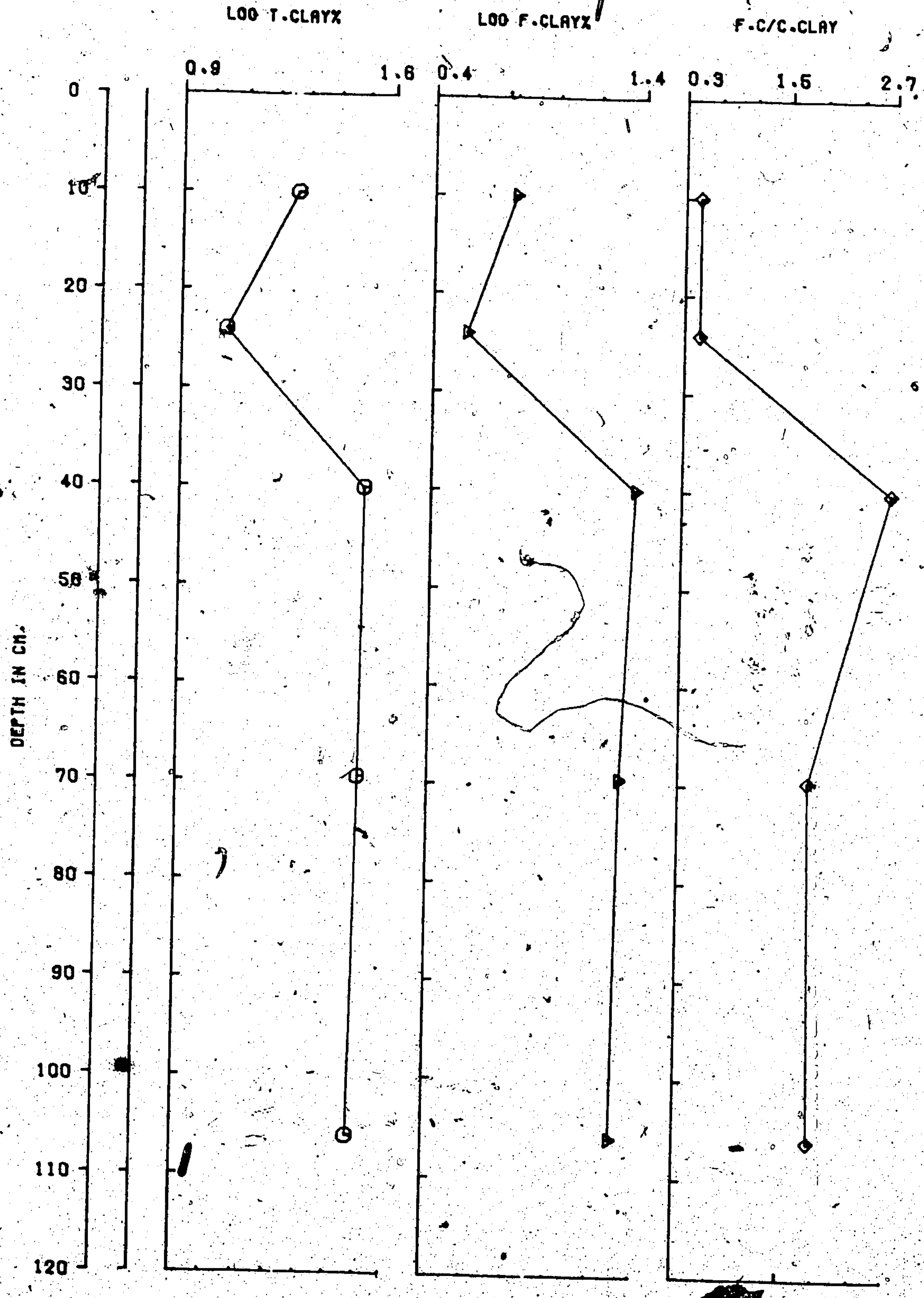


Figure 11. Total clay and fine clay content and fine clay:coarse clay ratio with depth for a Dark Gray Luvisol (site: 2)

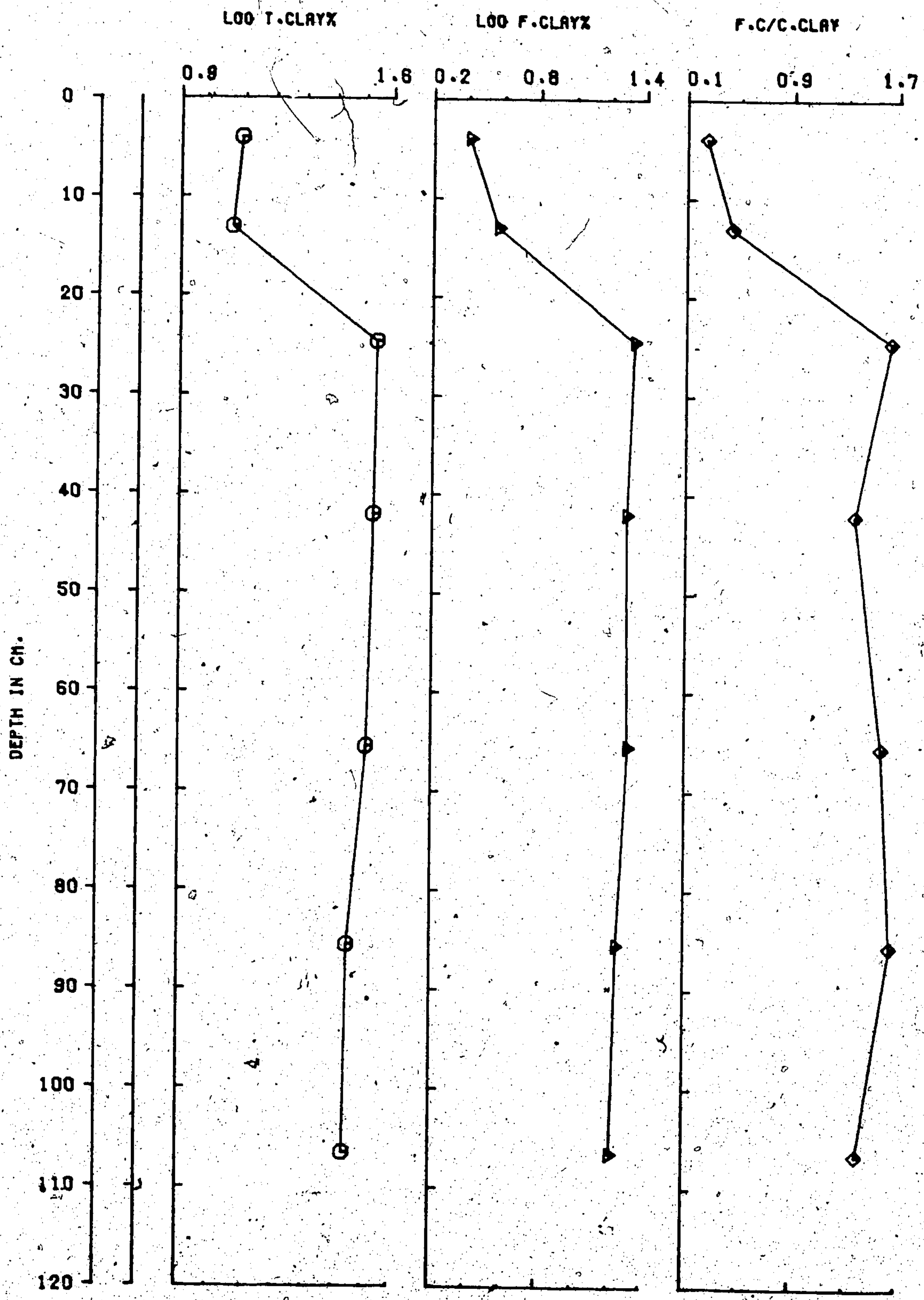


Figure 12. Total clay and fine clay content and fine clay:coarse clay ratio with depth for a Dark Gray Luvisol (site: 3).

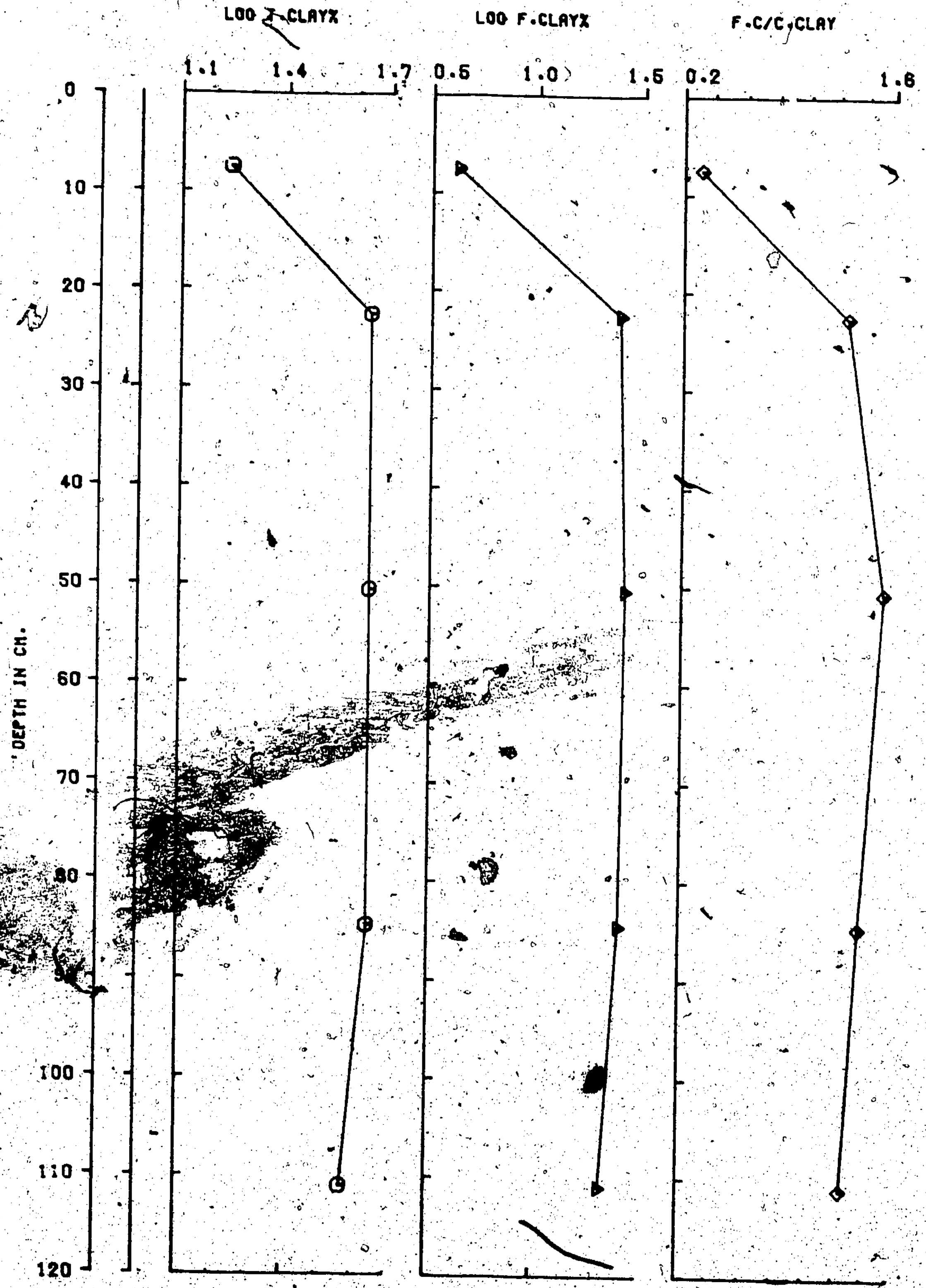


Figure 13. Total clay and Fine clay content and fine clay:coarse clay ratio with depth for an Orthic Gray Luvisol (site: 4)

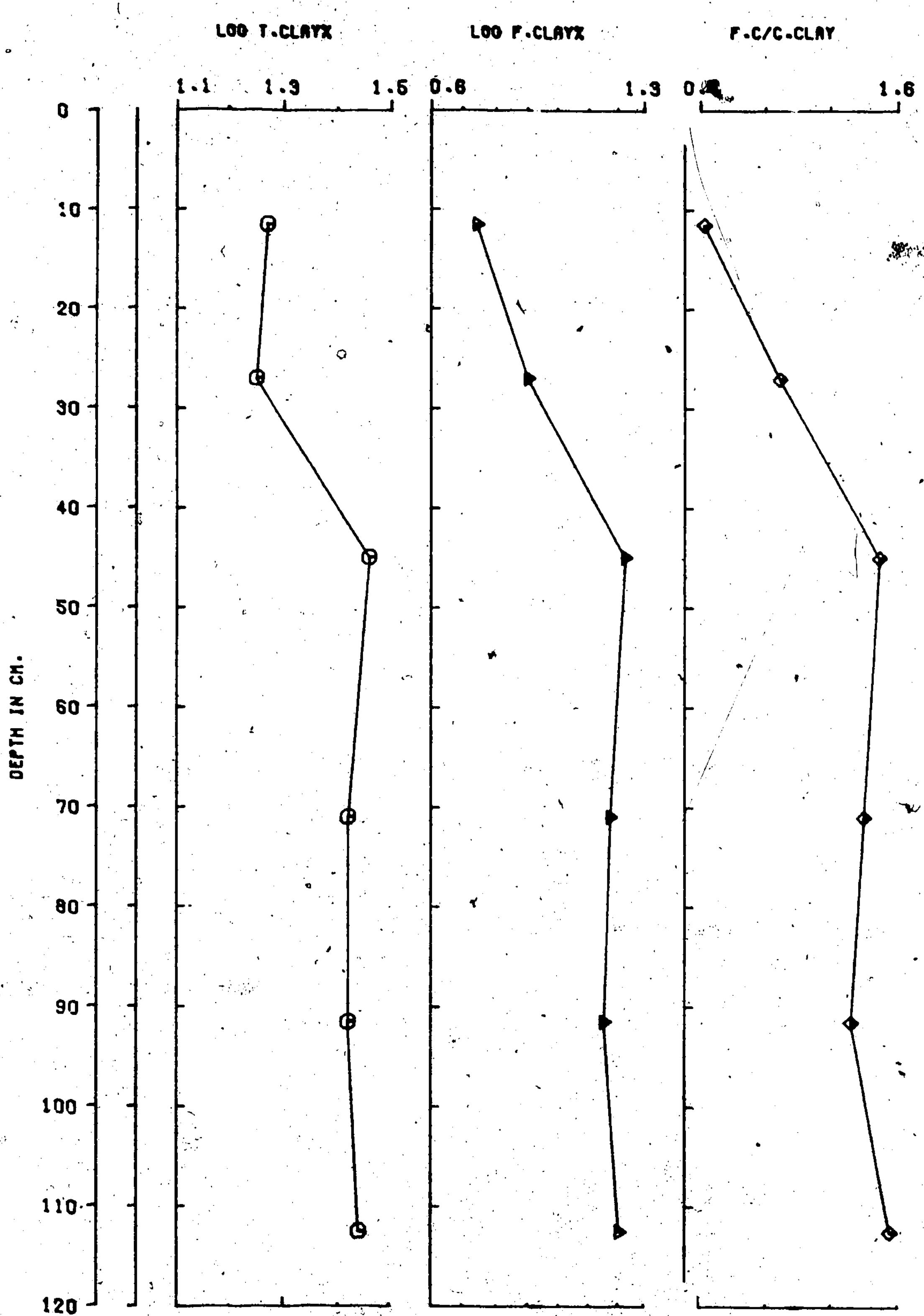


Figure 14. Total clay and fine clay content and fine clay:coarse clay ratio with depth for the Eluviated Black Chernozem (site: 5)..........

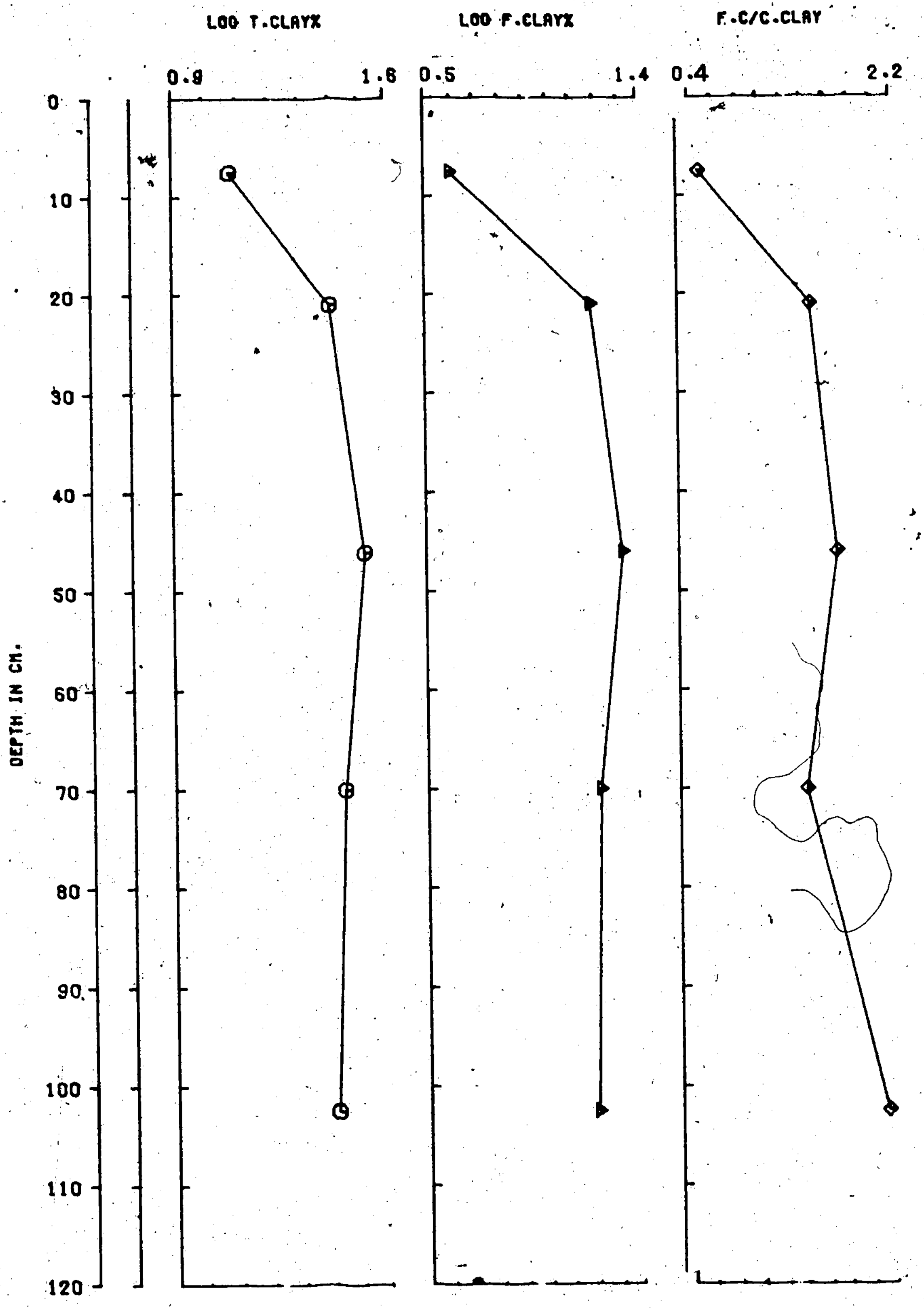


Figure 15. Total clay and fine clay content and fine clay:coarse clay ratio with depth for an Orthic Gray Luvisol (site: 6)..................

translocation. The range of fine clay to coarse clay ratios are 0.3 to 0.5 in the Ae and Ahe horizons, 1.4 to 2.7 in the upper Bt, 1.3 to 1.6 in the lower Bt and 0.9 to 2.1 in the C horizons (Tables 1 to 6, Appendix A). These values indicate an increase in the fine clay in the upper Bt horizons compared to that of either Ae or C horizons. However, the ratio of fine clay to coarse clay is highest in the C horizons for soils from sites 5 and 6 which further suggests illuviation of clay may extend beyond the B horizon. Fine clay to coarse clay ratios ranging from 0.1 to 0.8 for Ae horizons, 1 to 1.9 in Bt horizons and 0.8 to 1.2 in C horizons were reported for Luvisolic soils in Alberta (Pawluk, 1960b ; Lavkulich et al., 1964 ; McKeague et al., 1972; Twardy et al. (1973) reported that the continental tills of Alberta contained 11.8±1.8% fine clay and 23.4±2.3% total clay which gives fine clay to coarse clay ratios of 0.6 to 1.8.

The higher values for total clay, fine clay and fine clay to coarse clay ratios in the horizons below the Ae are evident for all soils examined in this study. In the Orthic Gray Luvisols, translocation of fine clay from the Ae and to a lesser extent AB horizons is evident (Figures 10 to 15). Except for site 1, distribution of total clay also suggests the illuviation process. At site 1, the fine clay content of the C horizon is 2% lower than the content of fine clay in the Bt1 horizon whereas coarse clay substantially increases by 7% in the C horizon. Consequently, total clay increases from 31% in the Bt horizon to 35% in the BCk and Cca horizons. These results may be a reflection of textural variations present in the parent till. The data also suggests clay migration from the Ah horizon of the Eluviated Black Chernozem occurred at least at some stage during soil development. The extent of this migration is comparable to that of the Ae horizons of both Orthic Gray and Dark Gray Luvisols. Formation of the thick Ah horizon found in the Eluviated Black Chernozem is probably a result of humification of a former Ae horizon. The former Ae horizon may have developed when the site was occupied by a dominance of forest vegetation after which grassland vegetation may have dominated the site resulting in humification of the previously formed Ae horizon. On the other hand, formation of Ahe horizons in Dark Gray Luvisols is thought to be a result of degradation of Ah horizons previously formed under grass vegetation. Degradation of former Ah horizons is thought to be a result of forest invasion onto areas of grass vegetation in the forest-prairie transition zone (Pettapiece, 1969).

Morphological observations in the field (Appendix A), micromorphological observations and particle size data all indicate the presence of illuvial layer lattice clay in Bt horizons. Mineralogical properties of profiles associated with clay translocation are expected to be affected by illuviation through dilution and negative enrichment of mineral consitutents due to both differential and/or non-differential translocation. The greatest effect attributable to illuviation is expected to occur in the clay fraction. However, dilution and/or negative enrichment of minerals in the sand and silt fractions make it necessary in most cases to recalculate the mineralogical results on a clay free basis in order to accurately assess observed mineralogical differences within and among profiles.

b. Calcium carbonate equivalent

Calcium and magnesium carbonate were present in the parent materials of all the soils studied. Carbonate translocation by solution in Chernozemic and Luvisolic soils developed on calcareous glacial till is one of the principal pedogenic processes (Pawluk, 1960b; Bourne and Whiteside, 1962; McKeague and St. Arnaud, 1969; St. Arnaud and Herbillon, 1973). Dissolution of carbonates and its subsequent leaching occurs primarily through the action of percolating water. The dissolution is greatly enhanced in the soil environment by the presence of organic acids released as a result of biological activity (Pawluk, 1960b; Kononova, 1977). Depletion of free carbonates from A and B horizons and subsequent leaching of basic cations will cause an increase in hydrogen ion activity in these horizons. The nature and extent of carbonate leaching and accumulation will be presented and briefly discussed.

In order to make comparisons between horizons of differing contents of carbonates more meaningful, the contents are presented as percentages of the carbonate-free soil materials (Table 1 and Figure 16).

The average content of carbonates in C horizons is 10.7% at sites 2 and 6, 7.6% at sites 1,3 and 5 and 3.4% at site 4. On the other hand, the average depths to carbonate accumulation in these profiles varies from 59 cm at sites 2 and 6, 83 cm at sites 1,3 ~~and 5~~ and 98 cm at site 4. The respective average thickness of Bt horizons in these groups of profiles is 31 cm, 50 cm and 68 cm. So, amounts of carbonates present in Ck or Cca horizons are in general, inversely proportional to solum thickness. A similar relationship was noted by McKeague et al. (1972) particularly in soils containing dolomite as a major carbonate form. Such an inverse relationship probably reflects effectiveness of leaching waters in mobilizing and transfering carbonates into the parent geological material.

Size distribution of carbonates in Cca horizons indicate their accumulation in either silt or clay fractions and in the silt fractions of BCk and Ck horizons (Table 1 and Figure 16). Silt-sized carbonates accounts for 44 to 66% of total soil carbonate in Cca horizons and 50 to 61% in BCk and Ck horizons while clay-sized carbonates accounts for 25 to 50% of total carbonates in Cca horizons and 26 to 32% in BCk and Ck horizons. Most lime accumulated in horizons of maxiumum carbonate accumulation in Borolls of North Dakota was also found to be finely divided (Redmond and McClelland, 1959). Clay-sized carbonates in soil are thought to partly

Table 1. Calcium and magnesium carbonate content of soil and size separates expressed on a carbonate free basis.

Soil type	Horizon*	$CaCO_3$ equivalent %**				Clay fraction***	
		Sand	Silt	Clay	Whole Soil	$CaCO_3$%	$MgCO_3$%
O.G. Luvisol	BC	ND	ND	ND	0.10	1.12	0.13
(site 1)	Cca	1.81	14.1	5.22	6.16	4.46	1.01
D.G. Luvisol	CCa1	1.79	16.7	17.6	10.4	8.86	3.00
(site 2)	Cca2	1.89	14.9	17.3	9.89	7.01	2.75
D.G. Luvisol	BC	ND	ND	ND	0.50	1.64	0.38
(site 3)	Cca	1.30	13.5	10.9	7.30	8.46	1.36
O.G. Luvisol (site 4)	Ck	1.82	5.66	2.58	3.41	2.23	0.65
E.B. Chernozem	BCk	1.64	14.6	7.75	6.50	6.72	1.11
(site 5)	Cca	2.21	17.5	13.8	9.41	15.7	1.01
O.G. Luvisol	BC	1.02	2.83	1.92	1.73	2.88	0.5
(site 6)	Cca	2.78	18.5	20.2	10.7	19.9	0.0

* Based on field observations

** Average of triplicate analyses

*** $CaCO_3$ and $MgCO_3$ equivalent calculated from CaO and MgO content of NH_4-saturated clays

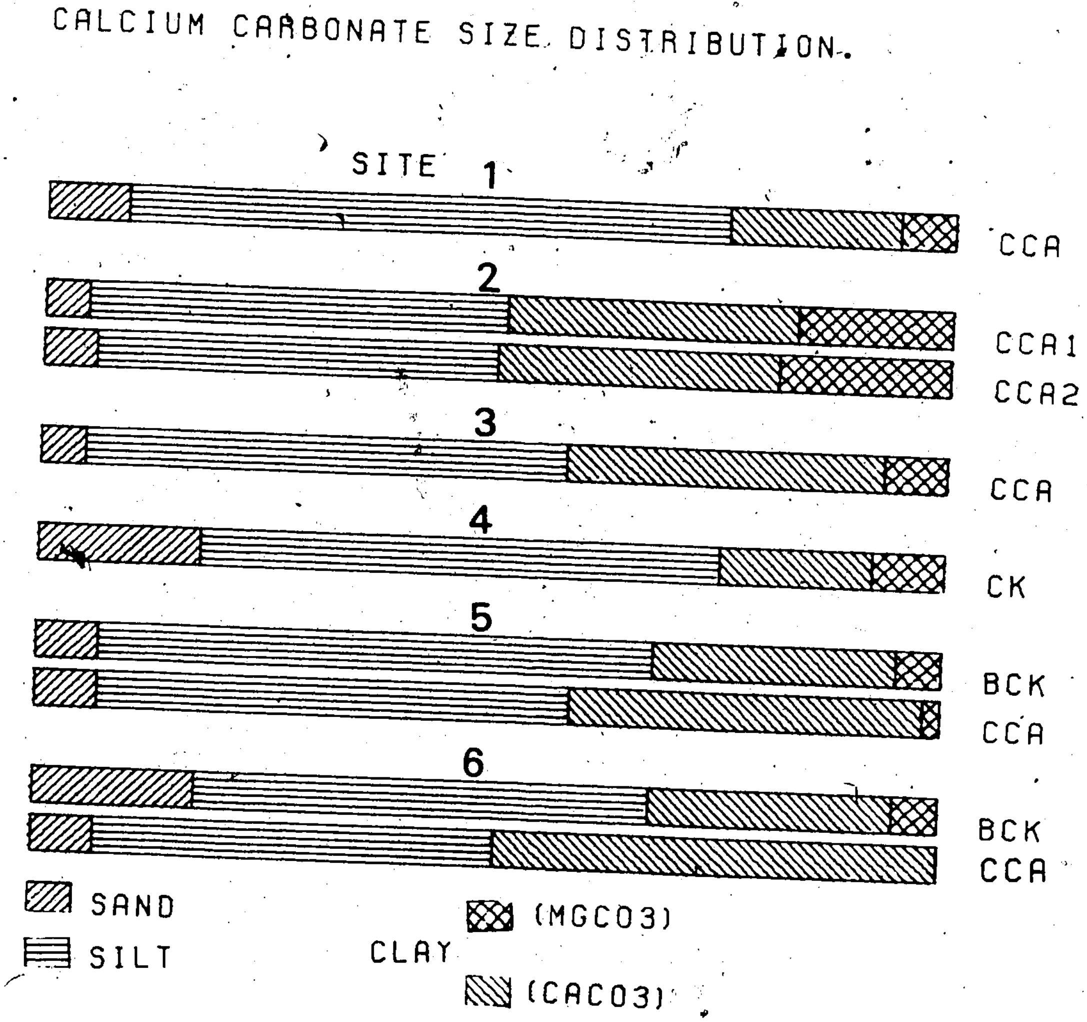


Figure 16. Bar-graphs showing distribution of carbonates in the various size fractions of the soils and the relative amounts of $CaCO_3$ and $MgCO_3$ in the clay fraction..........................

originate from dissolution of inherited limestone and/or dolomite within the developing solum followed by translocation and reprecipitation in C horizons as very tiny carbonate crystals (Rostad and St. Arnaud, 1970). The sand-sized carbonate contribution which is probably all of an inherited origin is only 5 to 9% in Cca horizons and 7 to 18% in BCK and Ck horizons. Estimates of composition of the clay-sized carbonate fractions (Table 1 and Figure 16) reveal that about 60 to 100% consists of $CaCO_3$ with generally higher relative quantities of $CaCO_3$ than $MgCO_3$ in clay fractions separated from Cca horizons as compared to BCk and Ck horizons. The results for carbonates in the clay fraction from soils from sites 5 and 6 suggest that as carbonates precipitate in Cca horizons, magnesium tends to be excluded from the resulting carbonate mineral as has been suggested by St. Arnaud and Herbillon (1970) for upland Chernozemic soils of central Saskatchewan.

c. Exchangeable cations, pH and total exchange capacity

Slightly acidic to acidic reactions were observed for all surface horizons (Tables 1 to 6, Appendix A). Lower pH values were recorded for Ae horizons of Orthic Gray Luvisols as compared to slightly higher values for Ah and Ahe horizons of the Eluviated Black Chernozem and Dark Gray Luvisols. Values for pH generally decrease down the solum and reach neutral to mildly alkaline values where carbonates occur.

Total exchange capacity ranges from a low of 2.7 me/100 g to a high of 27.6 me/100 g (Tables 1 to 6, Appendix A). The low values are associated with horizons leached of clay and/or organic matter. The Bt horizons of site 4 have an average total exchange capacity of 22.9 me/100 g and contain an average of 41.7% total clay, 24.7% fine clay and 1.5% organic matter. At sites 2,3 and 6 the average exchange capacity of Bt horizons is 19.5±2.0 me/100 g and they contain a[illegible]erage of 33.4±0.9% total clay, 21.2±1.7% fine clay and 1.2[illegible]0.2% organic matter. Bt horizons at sites 1 and 5 have an average exchange capacity of 15.2±1.1 me/100 g and contain 29.1±1.9% total clay, 16.4±2.2% fine clay and 1.1±0.2% organic matter. These results indicate that differences in the total exchange capacity among different soils or different horizons are mainly attributable to differences in amounts of total clay, fine clay and organic matter. Similar relationships between total cation exchange capacity of soil and content of organic matter, clay and fine clay are also evident for C horizons where samples of lower clay content have the lowest cation exchange capacity.

Exchangeable calcium occupies more than 70% of the exchange complex in the Ahe and Ah horizons of sites 3 and 5 (Table 2). Cycling of calcium by plants is probably responsible for replenishment of calcium in humified surface horizons of the Dark Gray Luvisols and the Eluviated Black Chernozem. The high amount of exchangeable magnesium in the Bt horizon of site 2 (64% as compared to an average of 33±6%

Table 2. Exchangeable cations expressed as ratios and as percentages of the total exchange capacity.

Horizon	Exchangeable cations as percentage of total cation exchange capacity				Exchangeable Cation ratios	
	Na⁺	K⁺	Ca⁺⁺	Mg⁺⁺	Ca⁺⁺:Mg⁺⁺	K⁺:Na⁺⁺
	<u>Dark Gray Luvisol (site 2)</u>					
Ahe	0.5	1.2	43.6	29.2	1.5	2.4
Ae	1.5	1.5	28.6	37.6	0.8	1.0
Bt	0.5	3.3	21.1	64.0	0.3	6.1
Cca	1.3	1.2	ND	ND	ND	0.9
	<u>Dark Gray Luvisol (site 3)</u>					
Ahe	0.1	4.2	72.4	10.9	6.6	64.0
Ae	0.2	5.2	59.9	13.2	4.5	36.0
AB	0.1	2.8	57.5	24.0	2.4	28.0
Bt1	0.1	1.9	60.2	28.4	2.1	18.5
Bt2	0.2	1.9	55.7	31.5	1.8	10.7
BCk	0.3	2.1	ND	ND	ND	6.2
CCa	0.4	2.5	ND	ND	ND	6.8
	<u>Eluviated Black Chernozemic (site 5)</u>					
Ah	0.1	4.0	72.7	20.1	3.6	54.5
Ae	0.3	2.0	57.9	32.9	1.8	6.0
AB	0.3	1.7	47.6	36.1	1.3	5.2
Bt1	0.8	2.0	50.4	39.1	1.2	2.6
BCk	1.0	2.4	ND	ND	ND	2.4
CCa	0.9	2.2	ND	ND	ND	2.4

Table 3. Exchangeable cations expressed as ratios and as percentages of the total exchange capacity.

Horizon	Exchangeable cations as percentage of total cation exchange capacity				Exchangeable Cation ratios	
	Na^+	K^+	Ca^{++}	Mg^{++}	$Ca^{++}:Mg^{++}$	$K^+:Na^{++}$
	<u>Orthic Gray Luvisol (site 1)</u>					
Ae	4.1	5.9	53.3	10.0	5.3	1.5
AB	1.0	4.4	56.2	19.7	2.9	4.5
Bt1	0.1	3.9	49.3	20.9	2.4	29.0
Bt2	0.1	3.6	50.0	21.4	2.3	20.0
BCk	0.2	3.3	ND	ND	ND	15.0
CCa	0.3	2.6	ND	ND	ND	10.0
	<u>Orthic Gray Luvisol (site 4)</u>					
Ae	0.4	3.8	67.2	22.0	3.1	9.0
AB	0.4	1.4	52.5	30.1	1.7	3.6
Bt1	0.6	1.4	52.3	31.3	1.7	2.3
Bt2	0.6	1.3	58.2	32.9	1.8	2.4
CK	0.6	1.4	ND	ND	ND	2.3
	<u>Orthic Gray Luvisol (site 6)</u>					
Ae	0.3	3.4	46.4	18.9	2.5	12.0
AB	0.4	2.1	55.6	25.7	2.2	5.8
Bt	0.4	1.0	62.6	29.4	2.1	2.3
BCk	0.5	0.9	ND	ND	ND	1.9
CCa	1.0	1.4	ND	ND	ND	1.4

at sites 2 and 5) is attributed to the relatively high quantity of $MgCO_3$ in the parent material (Figure 16 and Table 1). In the Orthic Gray Luvisols (Table 3), the level of calcium occupying the exchange complex in Ae horizons is lower than in surface horizons (Ah and Ahe) of one Dark Gray Luvisol (site 3) and the Eluviated Black Chernozemic (site 5). The relatively higher intensity of acid leaching in Orthic Gray Luvisols compared to the Dark Gray Luvisol and the Eluviated Black Chernozemic and the greater ability of grass vegetation to cycle calcium compared to forest vegetation may account for the differences. The two main exchangeable cations in the soils are calcium and magnesium where they occupy more than 80% of the exchange complex in most B horizons and at least 60% in some Ae horizons. Ratios of exchangeable calcium to magnesium for the profiles (Tables 2 and 3) illustrate the redistribution of these cations resulting from pedogenesis. These ratios are highest in surface horizons and generally decrease with profile depth. Similar trends in the ratio of exchangeable calcium to magnesium were reported for Luvisolic soils of Alberta where the average ratios for five profiles decreased from 4.5 in the Ae horizon to 3.4 in AB and 3.7 in Bt horizons (Pawluk, 1960b).

Exchangeable sodium and potassium are present in relatively low quantities usually amounting to less than 6% (Tables 2 and 3). Such low values are attributed to the ease of displacement and high mobility of these monovalent cations in the soil system. Exchangeable potassium is generally high in surface horizons while distribution of exchangeable sodium within soil profiles is somewhat erratic but generally higher in the lower parts of the solum as compared to the upper parts. The relatively high amount of exchangeable potassium in surface horizons may be attributed to the continuous supply of potassium from decomposition of forest and grass litter and partial dissolution of potassium-bearing minerals such as mica and feldspars. The relatively high content of exchangeable sodium in the lower portions of some profiles could be due to less effective leaching with soil depth or periodic influence of ground water.

d. Organic matter content

The decrease in organic matter content in A horizons proceeding from the Eluviated Black Chernozemic to the Orthic Gray Luvisols (Tables 1 to 6 , Appendix A) reflects the change in the nature of additions of organic residues. Where grass vegetation dominates (Eluviated Black Chernozemic), much of the organic debris is added into the soil and upon humification a mineral horizon highly enriched in organic matter forms. With forest vegetation, most of the organic debris added to the soil is in the form of leaf fall which accumulates on, rather than in, the soil. As a consequence, there is little incorporation of organic residues with mineral material in the Orthic Gray Luvisols. The Dark Gray Luvisols, associated with a mixed tree-shrub-grass vegetation, contain intermediate amounts of organic matter in their A horizons. Horizons of clay accumulation also display slight enrichment of organic matter in comparison to amounts in their respective underlying C horizons. This enrichment in B horizons is probably due to translocation of colloidal humus from surface horizons in response to leaching by percolating waters.

The content of organic matter in the clay fraction from horizons of organic matter accumulation comprise 13.1% and 18.5% in Ahe horizons of Dark Gray Luvisols (sites 2 and 3) and up to 31.6% in clay from the Ah horizon of the Eluviated Black Chernozemic which represents 19 to 26% of the total content of soil organic matter. Part of this material is probably associated with the clay fraction of these horizons through clay-organo complexing.

B. Mineralogical Characteristics of the Sand, Silt and Clay Fractions

Mineralogical characterization of coarse and medium sand (2-0.25 mm), fine sand (0.25-0.05 mm), coarse and medium silt (5-50 micrometer), fine silt (5-2 micrometer), clay (<2 micrometer) and fine clay (<0.2 micrometer) will be presented in this section.

a. Coarse and medium sand fraction

Soils contained less than 15% (weight basis) coarse and medium sand in those from sites 1,2 and 3 and up to 23% in soils from sites 5 and 6 (Tables 1 to 6, Appendix A).

Mineralogical characterization of the coarse and medium sand was facilitated by specific gravity separations into four density fractions, (<2.53, 2.53 to 2.59, 2.59 to 2.72 and >2.72 g/cc). The amounts of these specific gravity separates are listed in Tables 4 and 5. Petrographic examination indicated that the four separates were dominated by characteristic mineral assemblages.

The less than 2.53 g/cc separate consisted of weathered feldspar grains and coal fragments. Quantities of this separate remained relatively uniform within profiles and constituted less than 1.5% of the total coarse and medium sand fraction. The relatively high content of the <2.53 g/cc separate in some C horizons (Sites 1 and 2) may be related to the high abundance of coal fragments observed in these horizons during sampling and by binocular microscopic inspection. Weathered feldspar grains were observed in both the solum and in parent materials (C horizons) which suggests they are of an inherited origin rather than a result of pedogenesis. The surface morphologies of selected samples are illustrated in Plates 1 and 2. Solution pitting and alteration to secondary minerals and amorphous materials are the main features displayed by minerals in the <2.53 g/cc separate. The abundance of solution pitting on the surface of the feldspar grains (Plate 2) indicate moderate weathering has taken place. However, some areas appear to be

Table 4. Composition of the coarse and medium sand-sized fraction according to density separates. (sites 1,4 and 6)

Horizon	content (%)* <2.53 g/cc	2.53-2.59 g/cc	2.59-2.72 g/cc	>2.72 g/cc
	Orthic Gray Luvisol (site 1)			
Ae	0.5	9.6	88.5	1.4
AB	0.5	8.1	89.8	1.6
Bt1	0.6	11.1	86.6	1.7
Bt2	0.6	11.3	86.4	1.7
BCk	1.0	10.9	85.9	2.2
Cca	1.2	11.4	84.0	3.4
	Orthic Gray Luvisol (site 4)			
Ae	0.7	6.0	91.0	2.3
AB	0.7	6.1	91.5	1.7
Bt1	0.5	4.6	93.8	1.1
Bt2	0.7	4.7	93.5	1.1
CK	0.6	5.0	92.6	1.8
	Orthic Gray Luvisol (site 6)			
Ae	0.3	3.7	94.9	1.1
AB	0.6	3.3	95.5	0.6
Bt	0.2	3.5	95.5	0.8
BCk	0.4	3.4	94.7	1.5
Cca	0.4	3.4	94.0	2.2

*average of two replicates.

Table 5. Composition of the coarse and medium sand-sized fraction according to density separates. (sites 2,3 and 5)

	content (%)*			
Horizon	<2.53 g/cc	2.53-2.59 g/cc	2.59-2.72 g/cc	>2.72 g/cc
		Dark Gray Luvisol (site 2)		
Ahe	0.4	8.0	90.3	1.3
Ae	0.3	5.4	92.9	1.4
Bt	0.3	7.2	90.6	1.9
CCa1	0.6	11.2	86.0	2.2
CCa2	1.0	12.2	83.0	3.8
		Dark Gray Luvisol (site 3)		
Ahe	0.5	8.4	89.3	1.8
Ae	0.3	8.8	88.7	2.2
AB	0.5	8.9	89.3	1.3
Bt1	1.4	9.4	87.7	1.5
Bt2	1.3	9.8	87.0	1.9
BCk	0.7	10.7	85.8	2.8
CCa	0.5	9.9	86.1	3.5
		Eluviated Black Chernozemic (site 5)		
Ah	0.4	6.5	91.8	1.3
Ae	0.9	6.0	92.6	0.5
Bt1	0.5	5.7	93.1	0.7
Bt2	0.4	5.6	92.0	1.4
BCk	0.4	6.1	92.6	1.5
CCa	0.4	6.6	91.3	1.7

*average of two replicates.

Description of Plate 1

A: sub-rounded feldspar grain showing extensive solution pitting as well as physical abrasion.

B and C: A large solution pit with striated stacks of secondary mineral and amorphous materials. The platy and almost hexagonal crystals suggest kaolinite.

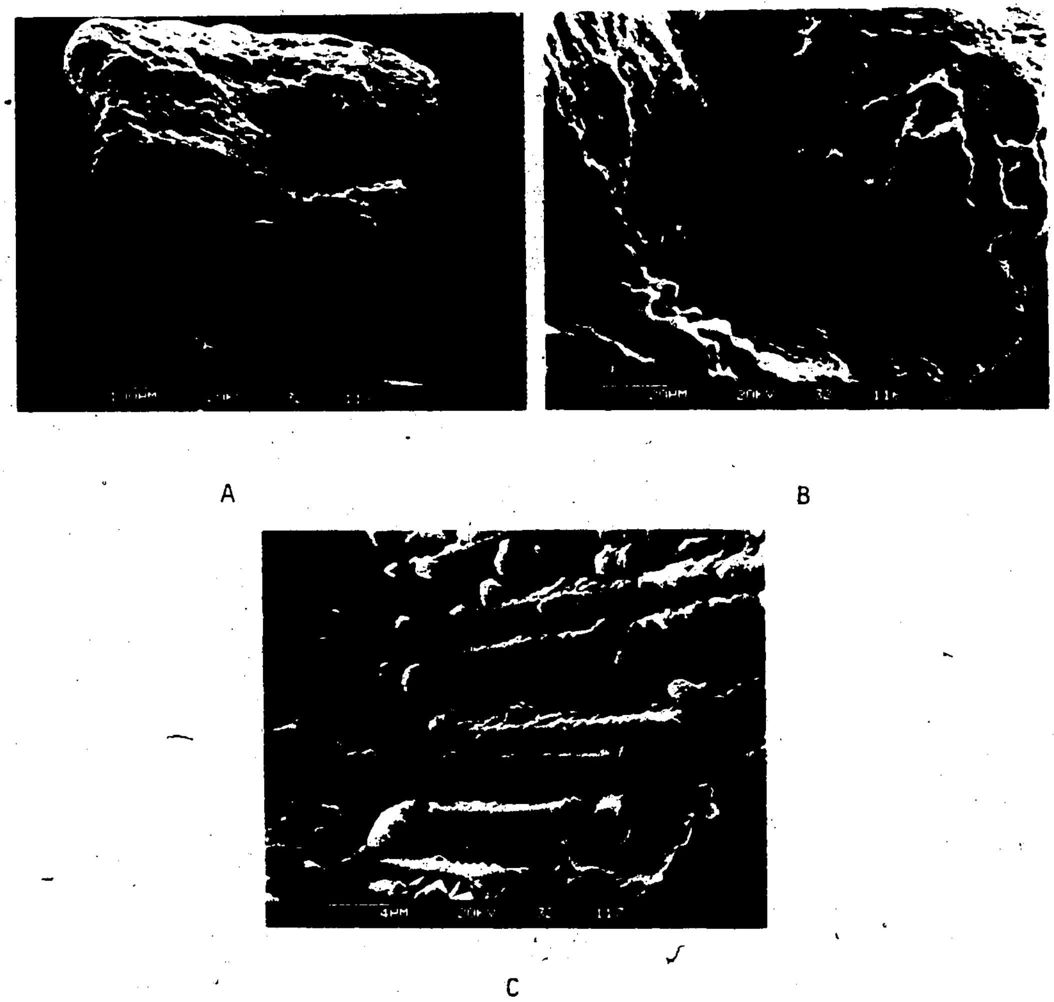


Plate 1. Scanning electron micrographs of a weathered feldspar grain separated from the Ae horizon of an Orthic Gray Luvisol (site 1).

Description of Plate 2

A to E: Micrographs A and E illustrate the appearance of moderately weathered feldspar grains. Micrograph C shows a sub-rounded grain. Solution pits are abundant on the surface of the grains (B and D). The flaky appearing crust on the surface of the feldspar grain (B) is suggestive of a secondary phyllosilicate, possibly kaolinite. The development of some regular fracture lines on the feldspar grain (E) is probably the result of frost action acting along a cleavage plane. The perforated appearance of grains suggests selective weathering has taken place.

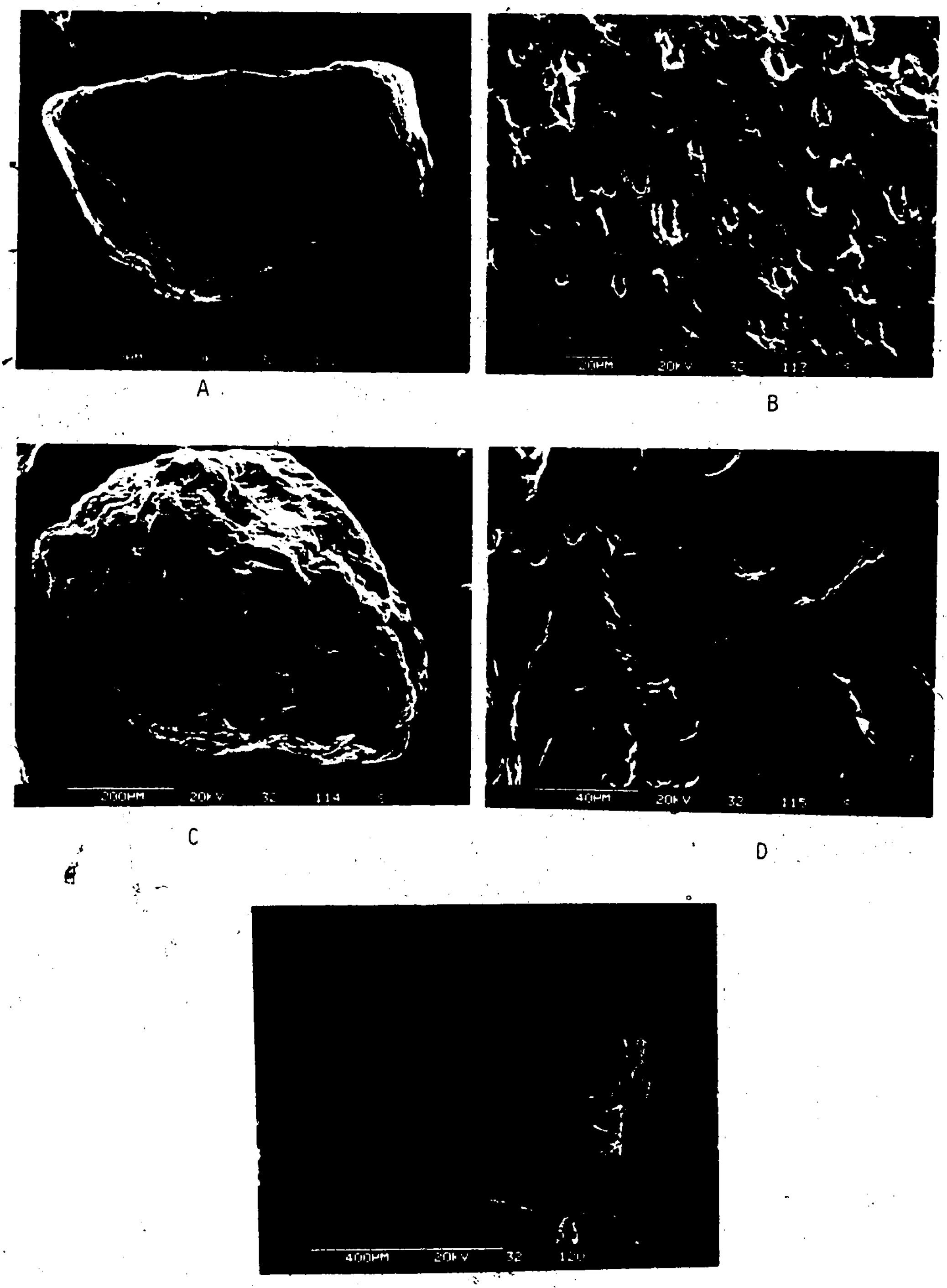


Plate 2. Scanning electron micrographs of weathered feldspar grains separated from the Ae horizon of the soil at site 1.

more resistant to weathering as indicated by the lack of solution pitting. This pitted mosaic appearance is possibly a reflection of differential weathering of feldspar members (or phases) which occur as perthitic intergrowths.

All density separates contained some rock fragments. Optical observations of rock fragments in the different specific gravity separates indicated that they consisted of minerals of fairly large crystals (0.1 to 0.4 mm). The mineralogical associations present in the rock fragments suggested an igneous origin although some rock fragments of metamorphic origin were also observed. Mineral associations observed in rock fragments in the 2.53-2.59 g/cc separate included mica (biotite), quartz and feldspars while associations of magnetite and quartz, feldspars and amphiboles were present in the heavy mineral separate (>2.72 g/cc).

The 2.53-2.59 g/cc separate constitutes less than about 12% (weight basis) of the total sand fraction (Tables 4 and 5) and contains mainly potassium feldspars (microcline, orthoclase and perthites). The feldspar grains are angular to sub-rounded and have a fresh appearance. A few feldspar grains displayed sericitization and other unidentifiable alteration products. These alteration products are probably a result of geological rather than pedological processes since all horizon samples appeared similar in thin section. Due to the presence of variable amounts of rock fragments in this specific gravity separate (2.53-2.59 g/cc) small variations in amounts between or among soil horizons can not be attributed to weathering. However, large variations exist in the quantity of potassium feldspars represented by the 2.53 to 2.59 g/cc specific gravity separates. Tills at sites 1,2 and 3 are fairly similar in potassium feldspar content whereas samples from sites 4 and 6, which are further from the Shield, contain approximately 45% of the amount found in samples from sites 1,2,3. The variations may be explained by mode of till deposition, direction of ice flow, dilution by local bedrock and possible variations in bedrock lithology of till source areas. Since sand grains in Cretaceous sandstones in Alberta seldom exceed 0.25 mm diameter (Pawluk and Bayrock, 1969) and siltstones and shales do not usually contain coarse particles, it would seem that coarse and medium sand is most probably of Shield origin. Percentages of the coarse and medium sand fraction are higher at site 6 than at sites 1,2 and 3 (an average of 19% compared to an average of 5%). In contrast, lower quantities of potassium feldspars are associated with site 6 rather than with sites 1,2 and 3 and therefore are probably not the result of dilution by sedimentary bedrock. This also appears to be the case at site 5 where amounts of K-feldspars and coarse and medium sand are comparable to those at site 6 despite the fact that site 5 is in the same general area as sites 1,2 and 3 (closer to the Shield area than site 6). Thus, one or more of the other factors mentioned above may be responsible for the lower K-feldspar content at sites 5 and 6 as compared to sites 1, 2 and 3 . At site 4, the average amount of coarse and medium sand is about 7% and thus dilution by bedrock may have contributed to the relatively low potassium feldspar content at this site compared to the other five sites.

The 2.59-2.72 g/cc separates are dominated by quartz and plagioclases. Although thin sections of this separate were examined, no attempt was made to estimate the amounts of plagioclases present since the number of feldspar grains were insufficient to obtain accurate data.

The amounts of heavy minerals (the >2.72 g/cc separate) are fairly similar for all sites although their distribution within profiles indicate the possiblity of weathering in A and B horizons(Tables 4 and 5). The amounts of heavy minerals separated using heavy liquids was very small which made it difficult to accurately determine the relative amounts of the mineral species present in this fraction. However, thin sections of heavy mineral separates were examined and the order of abundance appeared to be garnet > iron oxides > amphiboles > biotite. Garnet and in most cases magnetite were present as individual crystals while mica and amphiboles occurred in fragments associated with quartz and feldspars. Aggregates of iron oxides (possibly geothite) and quartz and reddish to reddish brown iron oxide concretions were also present in all horizons with relatively higher percentages in A and B horizons. Primary carbonates were present in separates from both BCk and Cca horizons. Carbonates occurred mainly in rock fragments as fine (10 to 50 micrometer) anhedral calcite crystals associated with some quartz.

b. Fine sand fraction.

This fraction was studied in greater detail than the medium and coarse sand since it constitutes the bulk of the sand fraction and has been suggested by Dreimanis and Vagners (1971) to approximate the "terminal grain size" of mineral comminution by glacial transport. Occurrance of relatively large and variable amounts of rock fragments in the various specific gravity separates of the coarse and medium sand fraction made it difficult to utilize the data to assess degree and nature of mineral weathering during soil development. Rock fragments in the fine sand fraction are present in relatively small amounts consequently their effect on interpreting the various analytical data should be minimal.

The fine sand was characterized following specific gravity separations. Data are reported as percentages of the total fine sand fraction (Tables 7 to 12, appendix A).

The content of the less than 2.53 g/cc specific gravity separate ranges from 0.2 to 1.7% in all samples. Scanning electron microscopic and petrographic examinations indicate that this portion is dominated by weathered minerals, especially feldspars (Plates 3 to 5). Pseudomorphic alteration of primary feldspar minerals is evident from the retention of their morphological structures. Secondary minerals such as kaolinite and possibly halloysite and amorphous materials are present as alteration products of feldspars. Electron micrographs of pitted feldspar grains similar to those presented in this study were obtained for deeply weathered granite in the Malaysia Peninsula (Eswaran and Bin, 1978).

The 2.53 to 2.59 g/cc and 2.59 to 2.72 g/cc specific gravity separates are dominated by potassium feldspars and quartz and plagioclase, respectively. These separates were analysed for K20, Na20 and CaO content and the mole per cent of KAlSi308 (Or), NaAlSi308 (Ab) and CaAl2Si208 (An) was calculated based on these oxide values and then illustrated in the ternary diagrams (Figures 17 to 19). The 2.53-2.59 g/cc fraction appears to be dominated by potassium feldspars and the 2.59 to 2.72 g/cc by feldspar having a bulk composition of oligoclase (Barth, 1969). Although these results give an average composition of feldspars present in the soils, it is difficult to ascertain the presence of specific alkali or plagiocase feldspars without additional information about the structural aspects of the mineral grains and the nature of admixtures and exsolution phenomena. For example, ternary diagrams indicate the dominance of potassium feldspars in the 2.53-2.59 g/cc separate but do not indicate whether the polymorphs are

Description of Plate 3

Micrographs A and B show differential weathering of the feldspar grain surface as seen by a scalloped appearance. The surface of a weathered face (micrograph B) shows differential weathering and the formation of secondary minerals and/or amorphous materials. The weathered portions are covered with tiny sheet-like bundles (particularly evident in micrograph A, upper left of the frame).

C: Altered feldspar surface displaying a flaky appearance. Edges of flakes appear rolled or curled and could represent halloysite or amorphous material.

D: Surface of a weathered feldspar grain which displays a variety of shapes ranging from almost cubic forms to thin platy structures and platy aggregates. Microprobe analysis would be necessary to identify the nature of these alteration products.

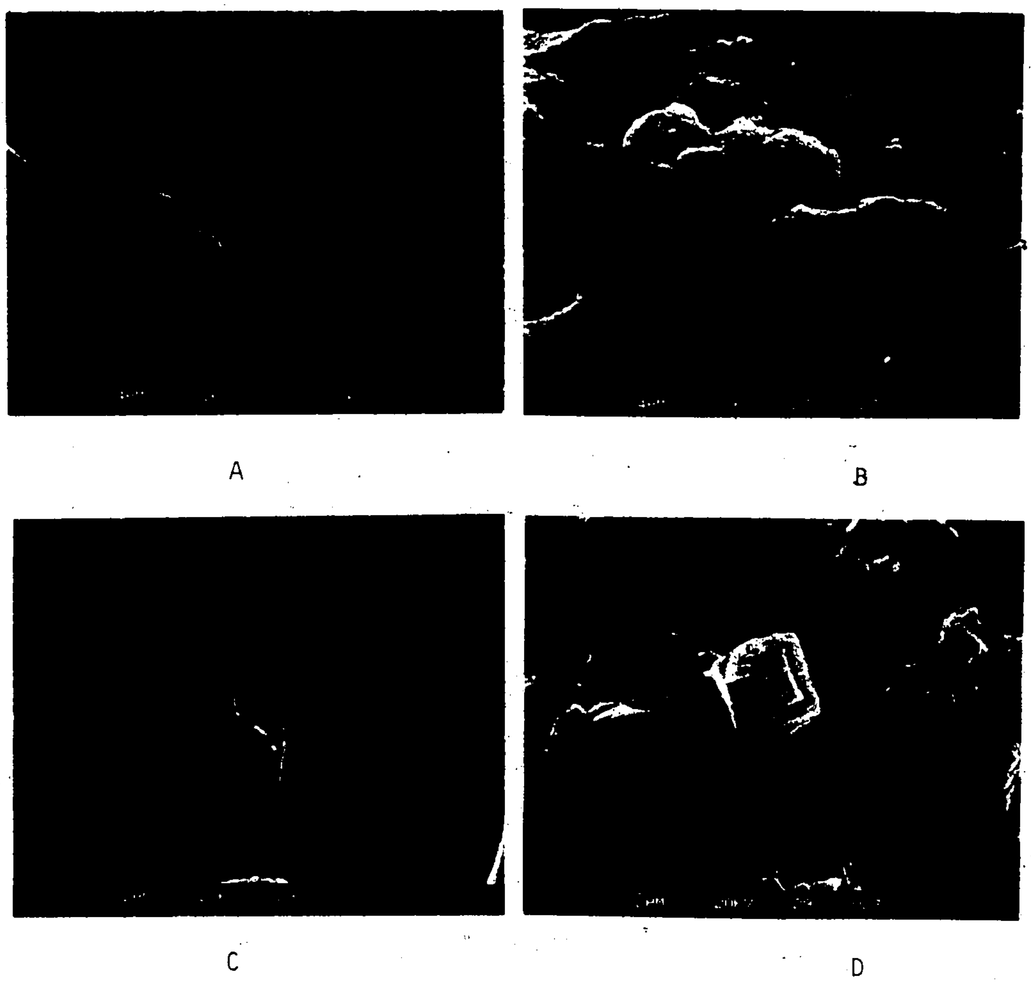


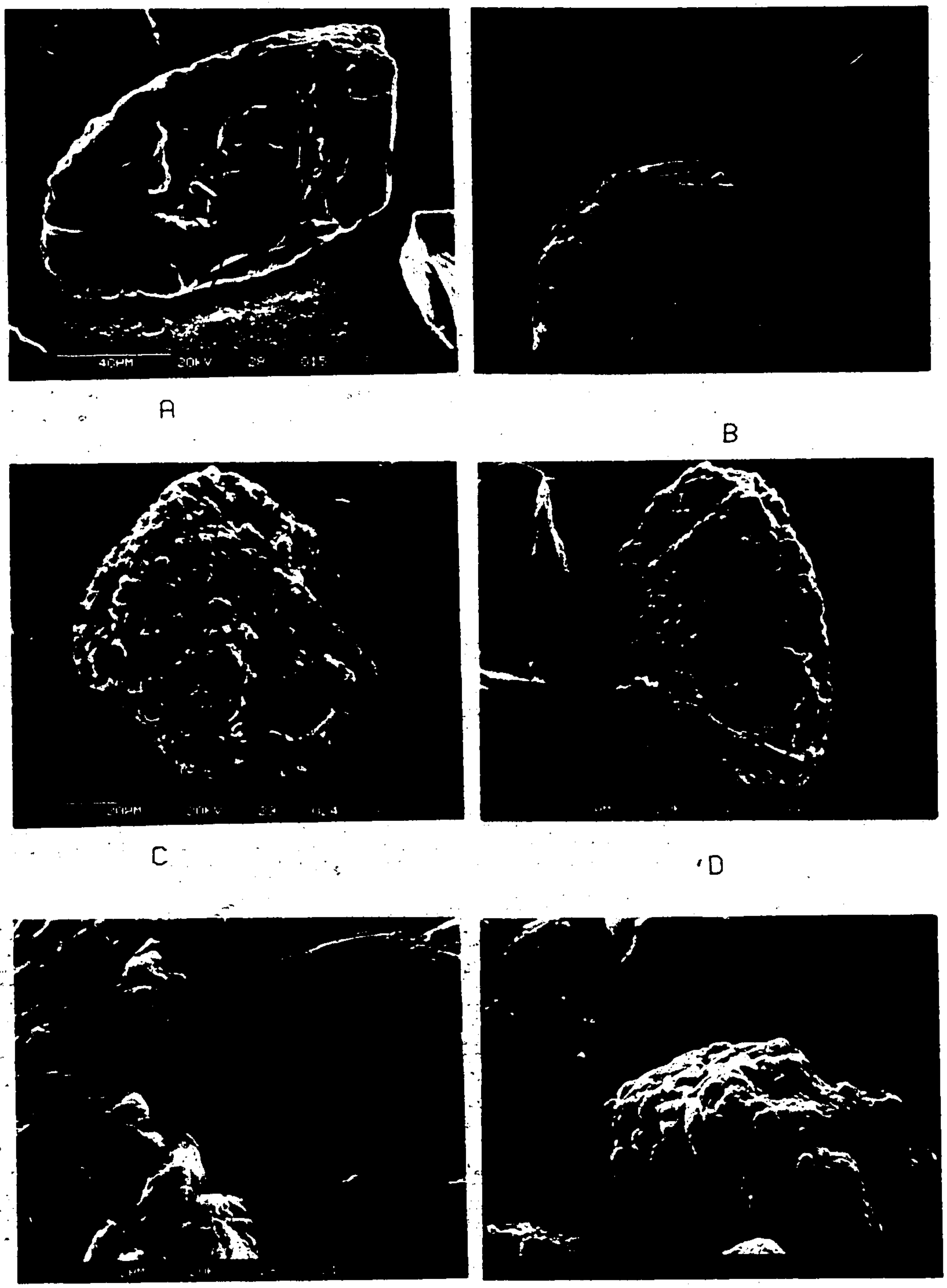
Plate 3. Scanning electron micrographs of weathered fine sand-sized feldspar grains separated from Cca horizons from two Orthic Gray Luvisols (A and B, site 1; C and D, site 6).

Description of Plate 4

A: Feldspar grain showing dissolution pits. Sides of dissolution pits are often parallel to the one good cleavage face. Pitting appears to develop in certain areas of the grain only, possibly indicative of different feldspar phases or selective weathering initiated by stress points or areas of inclusion of impurities.

B: The grain in the upper right corner shows the typical feldspar cleavage in two directions with no apparent pitting of the surface. These are probably freshly exposed faces. The second grain shows extensive solution pitting and formation of secondary mineral material.

C to F: Formation of kaolinite on feldspars as evidenced by numerous hexagonal shaped flakes and stacks of hexagonal and/or sub-hexagonal crystals. Stacks of platy crystals are most obvious in micrograph D.



A

B

C

D

E

F

Plate 4. Scanning electron micrographs of feldspar grains from the fine sand fraction from the Ae horizons of Orthic Gray Luvisols (A, D and E, site 6; B and F, site 1) and from a Cca horizon of an Orthic Gray Luvisol (C, site 6).

Description of Plate 5

A to D: Micrographs of material of biological origin. Originally, these grains were thought to represent some mineral substance. It is likely, however, these grains are diatoms.

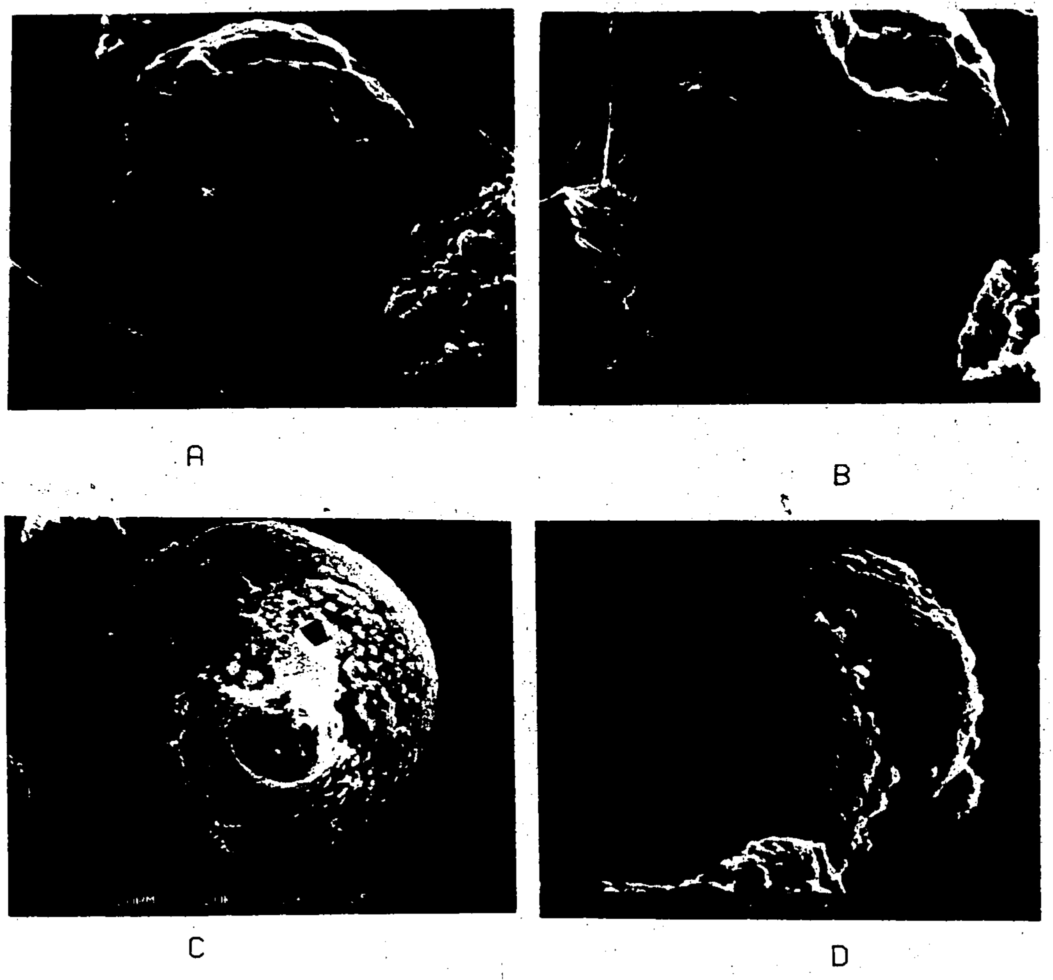


Plate 5. Scanning electron micrographs of sand-sized mineral material of unusual form separated from the Cca horizon from an Orthic Gray Luvisol (site 6).

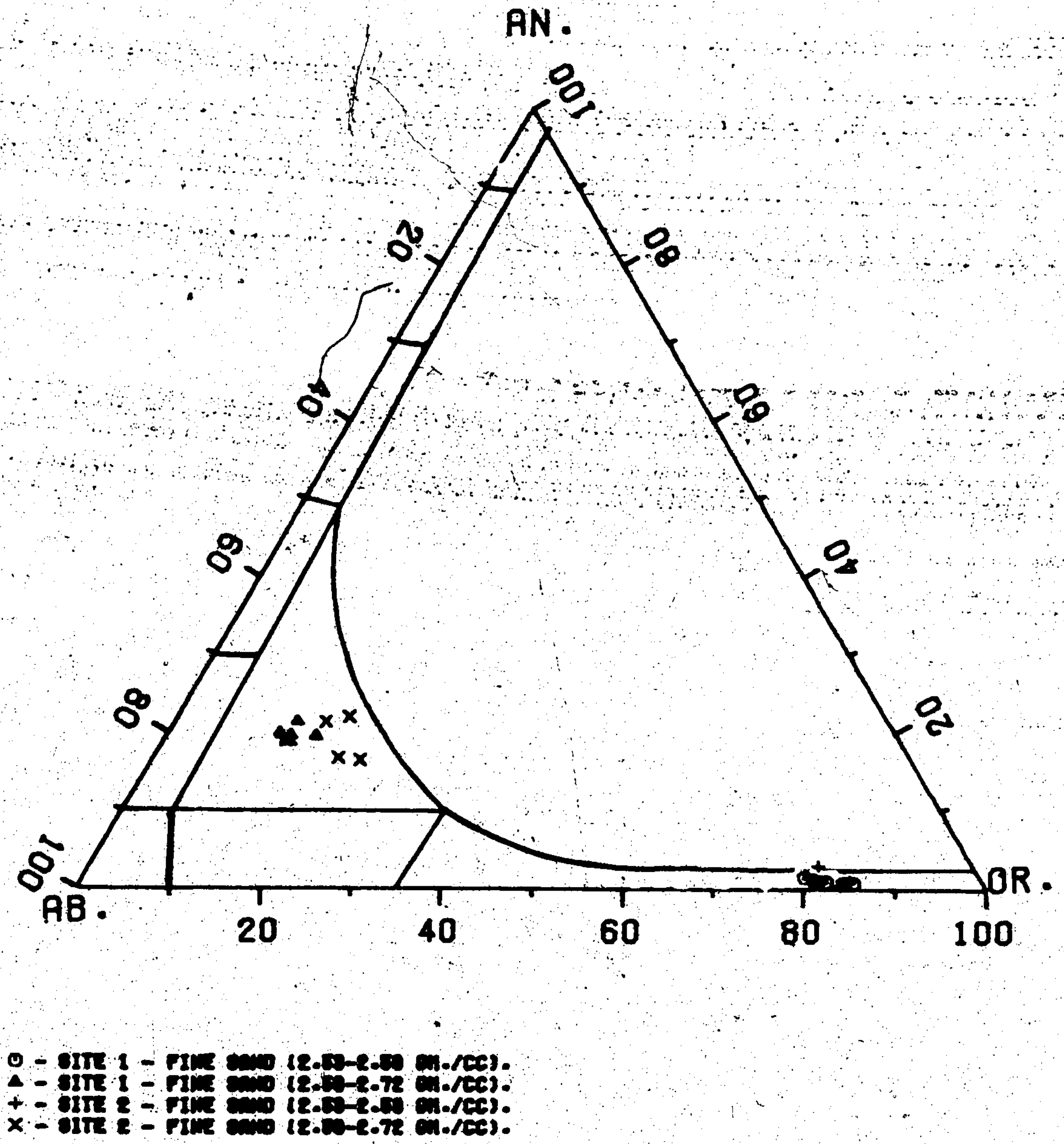


Figure 17. Composition of feldspars separated from soil horizons from soils at site 1 and 2.

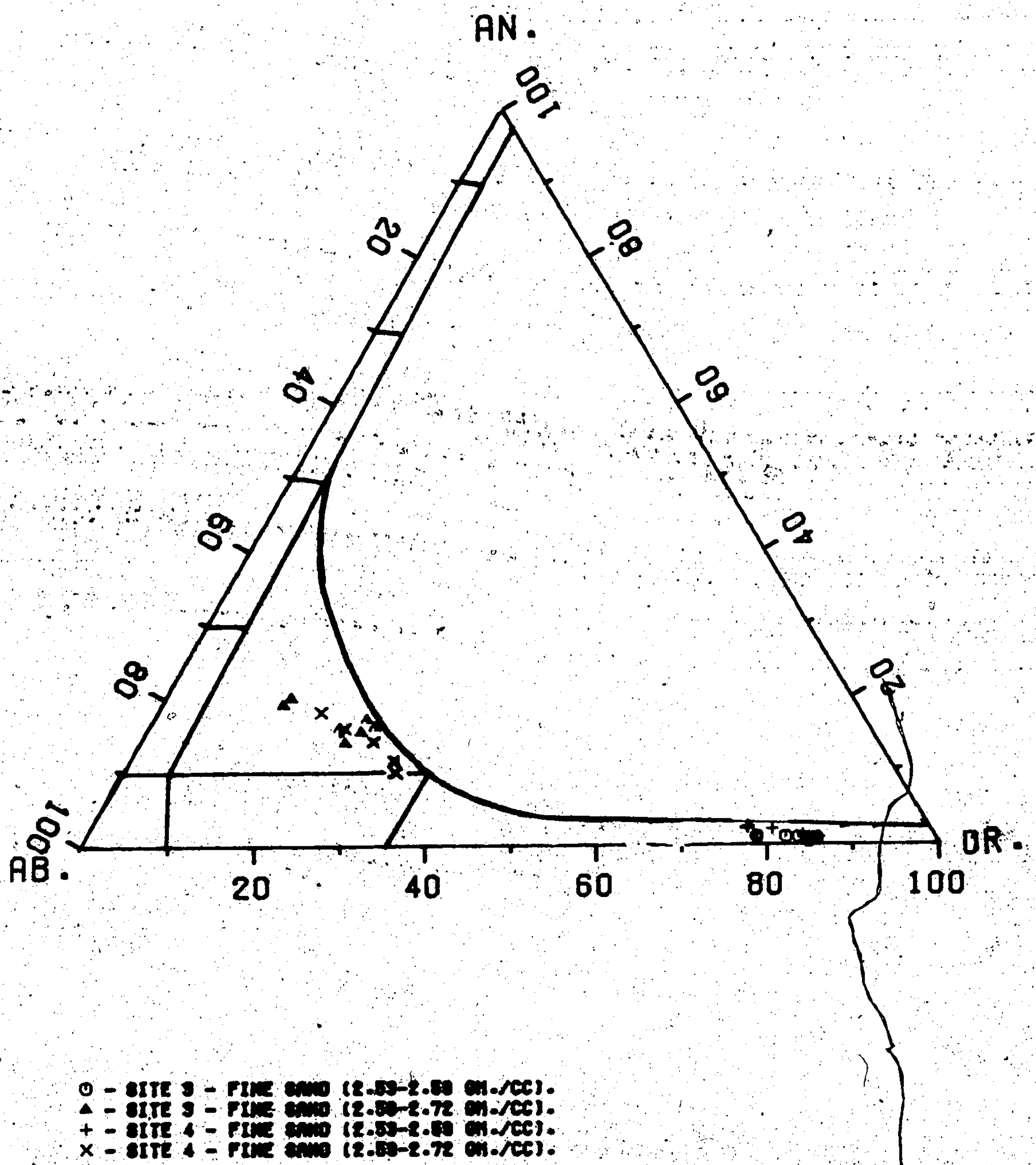


Figure 18. Composition of feldspars separated from soil horizons from soils at site 3 and 4.

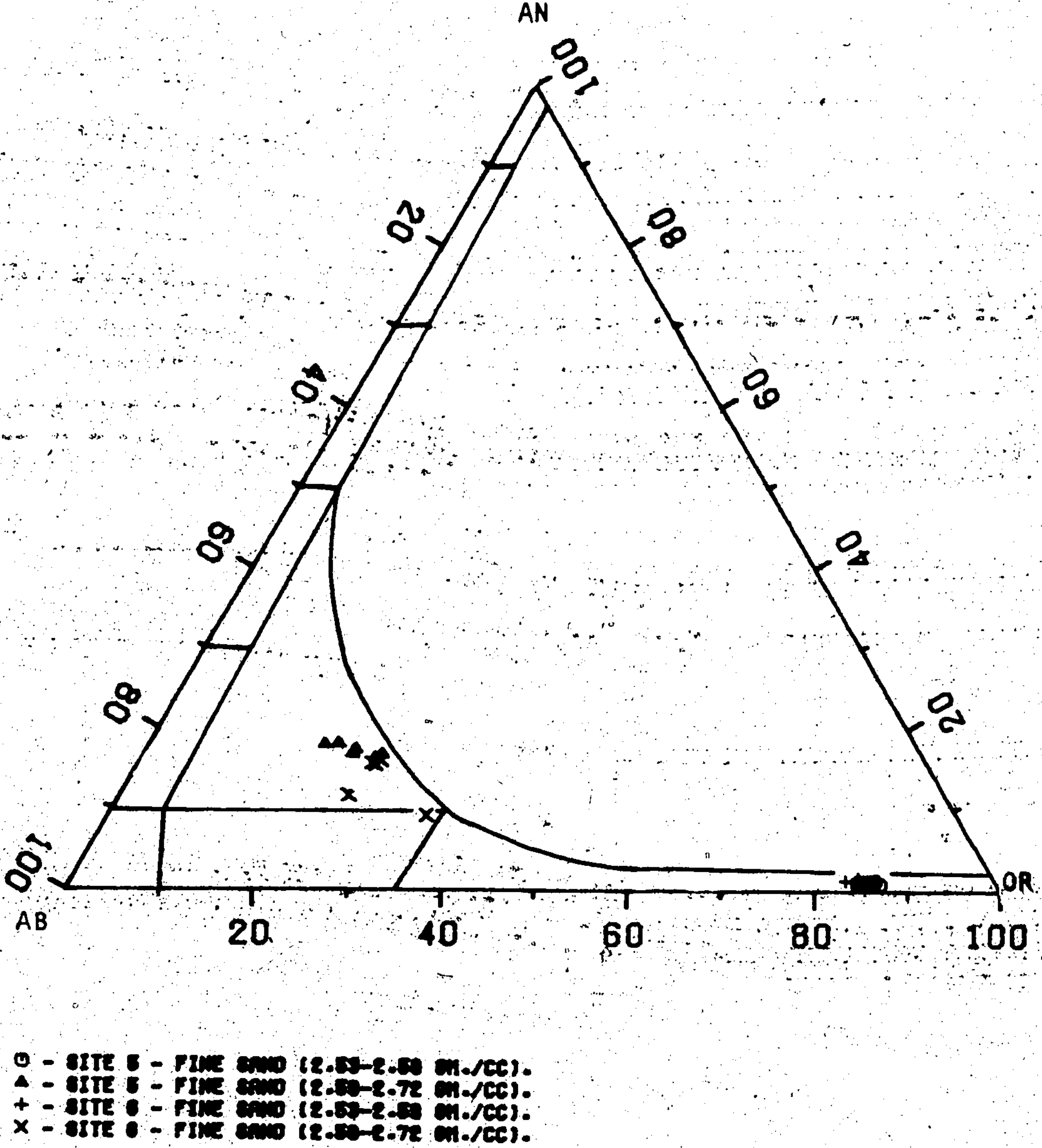


Figure 19. Composition of feldspars separated from soil horizons from soils at site 5 and 6.

orthoclase, microcline, perthitic or some combination of all of these minerals. Similarly, the diagrams (Figures 17 to 19) suggest the presence of oligoclase in the 2.59-2.72 g/cc separate but do not give information as to whether the bulk chemical composition is a result of an admixture of several feldspar species or not. Thus, in addition to bulk chemical composition, more detailed mineralogical investigations are necessary in order to fully characterize the nature of feldspars in the soils of this investigation.

Further characterization of feldspar minerals was accomplished utilizing X-ray diffraction methods, petrographic examinations and scanning electron microsopic observations. These methods provided a means of assessing the specific feldspar species present and obtaining information relating to the structural state of feldspars (polymorphic forms).

Tables of calculated spacings and intensities by Borg and D.K. Smith (J.V. Smith, 1974) and simulated X-ray powder patterns of feldspar minerals (Murfurin, 1966) were used to index the X-ray data obtained in this study. Data on Miller indices, peak positions and intensity empirically derived from measurement of nearly 100 X-ray diffraction patterns of natural alkali feldspars were also used (Wright and Stewart, 1968). Other sources of X-ray diffraction behavior of feldspars included the data published by the Joint Committee on Powder Diffraction Standards (1974).

Diffraction patterns of pure feldspar species, an artificial mixture of feldspar species and also a mixture of feldspars with quartz, as illustrated in figure 20, were used to aid in identification of feldspar species. The figure illustrates several X-ray diffraction properties of mixtures of feldspars such as:

(1) the difficulty in differentiating all spacings of K-feldspars from those of plagioclase. The most reliable peak to differentiate these feldspars is the -201 reflection which for K-feldspars occurs at about 20.8° 2θ(4.27 Å) and for plagioclases at about 21.9° 2θ(4.06 Å). Other peaks that may be useful in verifying the presence of K-feldspars admixed with plagioclases are:

a. presence of the 111 reflection of K-feldspars at about 22.2° 2θ(4.00 Å).

b. splitting of the peak at 25.4° 2θ into the -1-12 reflection of K-feldspars at about 25.5° 2θ(3.49 Å) and the -1-12 reflection of albite at about 25.3° 2θ(3.52 Å).

c. presence of the 002 reflection of K-feldspars at 27.5° 2θ(3.24 Å). The peak intensity is very high and may be confused with the 002 reflection of albite at 27.8° 2θ(3.21 Å).

d. presence of the 131 reflection of microcline at about 29.4° 2θ(3.04 Å) or the 131 reflection of orthoclase at about 29.8° 2θ(3.00 Å) (Figure 20, blackened peak).

Description of Figure 20

X-ray diffraction patterns showing reflections for different feldspar species either alone or admixed. Symbols indicate the diffraction lines of single mineral constituents. Numbers and their heights indicate peak positions and relative intensities for albite and labradorite. The solid line represents the diffraction of microcline and the dotted line indicates the X-ray pattern when microcline is mixed with albite and oligoclase in a 1:1:1 proportion by weight with trace amounts of quartz. Miller indices are shown for various peaks and are designated by Ab. for albite, Mi for microcline, and the rest will be either for microcline and/or orthoclase.

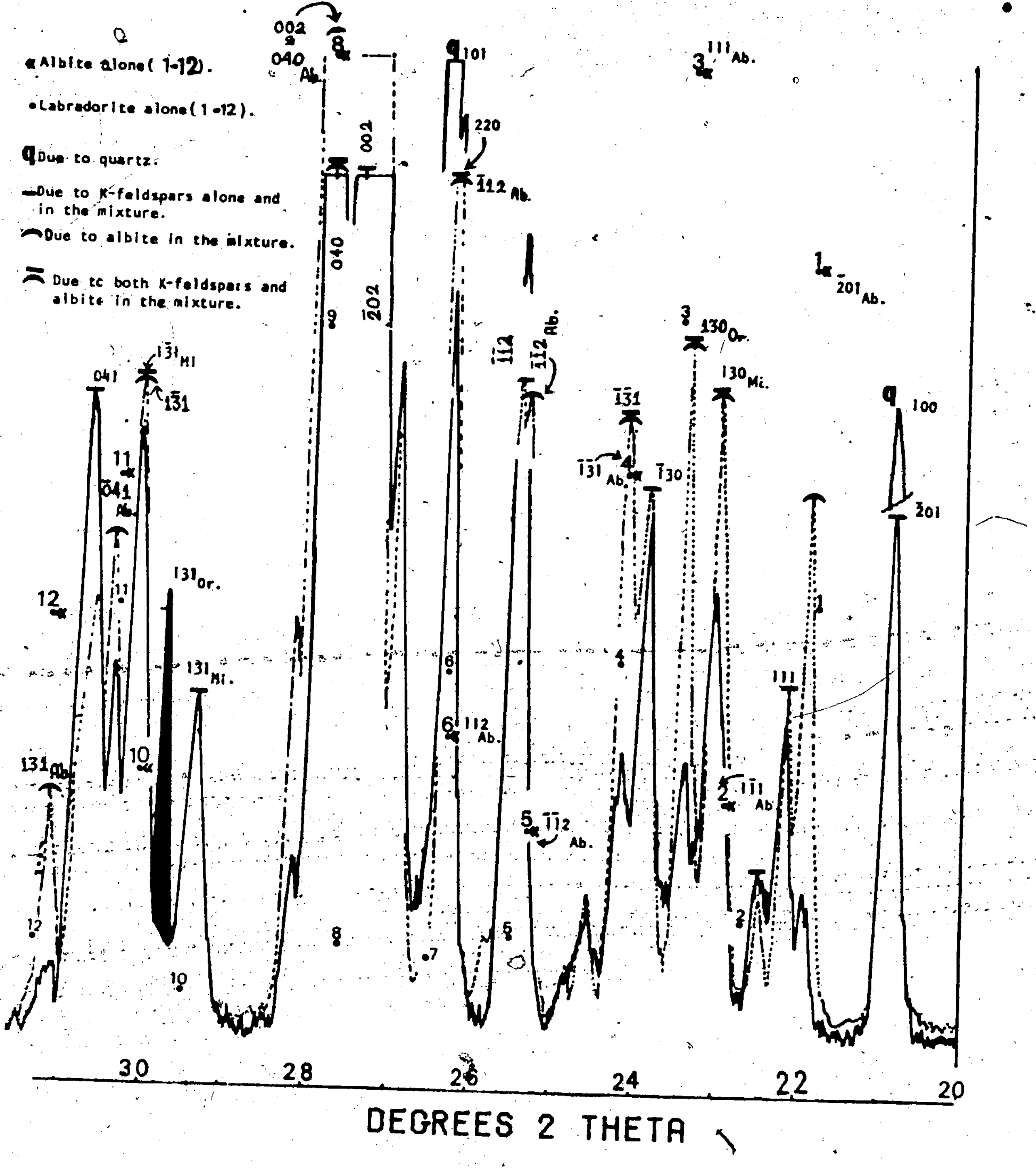


Figure 20. X-ray diffraction patterns showing feldspar species either alone or admixed (see preceeding page for description)..............

e. presence of the 041 reflection of K-feldspars at about 30.7° 2θ(2.91 Å).

(2) if quartz is present in the mixture it can easily be detected in the presence of K-feldspars by the broadening of the top of the -201 reflection of K-feldspars at about 20.8° 2θ(4.27 Å) towards the 100 reflection of quartz at 20.85° 2θ(4.26 Å >. Detection of quartz and albite is accomplished by the 100 reflection of quartz at 20.85° 2θ and the -201 reflection of albite at 21.9° 2θ(4.06 Å). The 101 reflection of quartz at 26.65° 2θ(3.35 Å) can be distinguished from the 220 reflection of both K-feldspars and albite at 26.3° 2θ(3.39 Å).

(3) the presence of calcic feldspar (e.g. labradorite) when mixed with albite is difficult to ascertain as it may causes broadening of the albite peaks rather than producing descrete peaks of it's own (Figure 20). The presence of double peaks near 21.8° 2θ(4.08 Å) ,22.9° 2θ(3.88 Å),25.3° 2θ (3.52 Å) and 26.5° 2θ(3.36 Å) indicate the presence of more than one plagioclase feldspar phase.

X-ray diffractograms of the 50 to 100 micrometer sand and the 5 to 50 micrometer silt size fractions from two samples are presented to illustrate the use of X-ray data in identification of potassium feldspars and their structural states in the 2.53 to 2.59 g/cc separates (Figure 21 and Tables 6 and 7). The -201 reflection indicates the presence of potassium feldspars. Other reflections (111,002,041) from the samples (Tables 6 and 7) also confirm the presence of

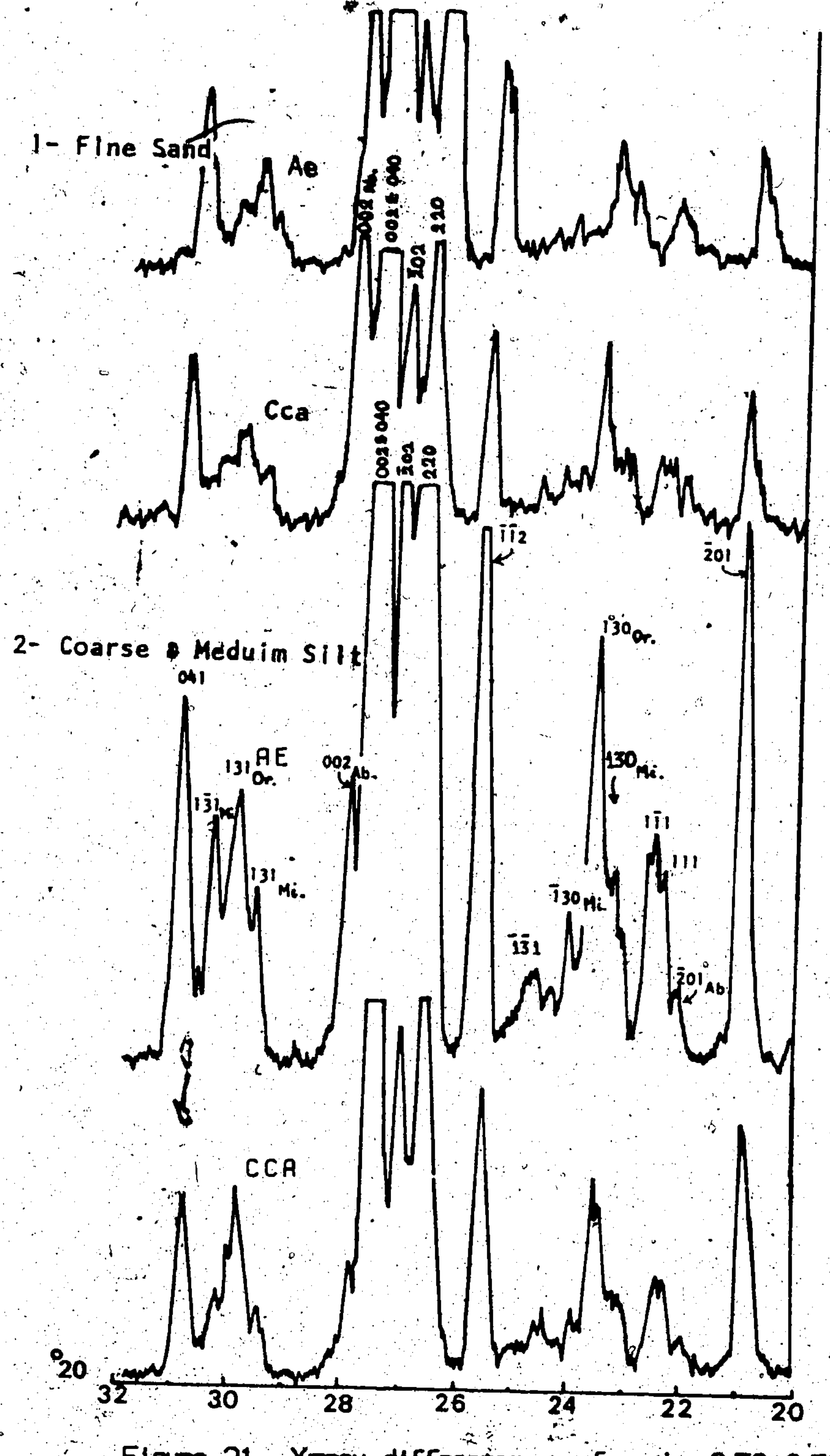


Figure 21. X-ray diffractograms for the 2.53-2.59 g/cc specific gravity separates from the Ae and Cca horizons of an Orthic Gray Luvisol soil (site 1) (all indices except as noted are for K-feldspars)..............................

Table 6. Bulk chemical composition and X-ray diffraction spacings of the 2.53 to 2.59 g/cc specific gravity separate of the fine sand fraction from the Ae and Cca horizons of an Orthic Gray Luvisol (site 1).

Horizon	Ae			Cca		
Bulk Composition	Ab15 Or84 An1			Ab18 Or81 An1		
X-ray diffraction peaks						
Miller Index	°2θ CuK$_\alpha$	d=Å	I*	°2θ CuK$_\alpha$	d=Å	I*
-201	21.03	4.224	20	21.03	4.224	21
-201 Ab	22.03	4.035	8	22.07	4.034	11
111	22.38	3.972	9	22.31	3.985	14
1-11	22.48	3.955	12			
130 Mi	23.23	3.829	14			
130 Or	23.58	3.773	19			
-130 Mi	23.73	3.749	12	23.98	3.711	13
-1-31	24.30	3.663	11	24.28	3.666	12
-1-12	25.63	3.476	27	25.68	3.469	11
220	26.63	3.347	53	25.60	3.351	27
-202	27.08	3.293	47	27.08	3.293	48
002, 040	27.48	3.246	100	27.50	3.243	100
002 Ab	27.93	3.194	45	27.83	3.206	6
131 Mi	29.52	3.026	10	29.52	3.026	10
131 Or	29.80	2.998	17	29.83	2.995	15
1-31 Mi	30.23	2.956	11	30.28	2.957	6
041	30.80	2.903	23	30.80	2.903	23

* I = relative intensity, calculated from X-ray patterns by assingning 100% intensity to the highest peak.

Table 7. Bulk chemical composition and X-ray diffraction spacings of the 2.53-2.59 g/cc specific gravity separate of the coarse and medium silt fraction from the Ae and Cca horizons of an Orthic Gray Luvisol (site 1).

Horizon	Ae			Cca		
Bulk Composition	Ab15 Or84 An1			Ab13 Or86 An1		
X-ray diffraction peaks						
Miller Index	°2θ CuK$_{\alpha}$	d=Å	I*	°2θ CuK$_{\alpha}$	d=Å	I*
-201	21.03	4.224	57	21.07	4.216	28
-201 Ab	22.12	4.019	13	22.12	4.019	13
111	22.38	3.972	24	22.47	3.963	13
1-11	22.52	3.948	28	22.59	3.936	13
130 Mi	23.22	3.831	24	23.29	3.819	6
130 Or	23.57	3.774	47	23.62	3.767	24
-130 Mi	24.02	3.705	20	24.07	3.697	10
-1-31	24.32	3.660	12			
-1-12	25.57	3.484	14	25.62	3.477	10
220	26.67	3.342	90	26.69	3.340	50
-202	27.07	3.294	79	27.12	3.288	37
002, 040	27.52	3.241	100	27.57	3.235	100
002 Ab	27.97	3.190	32	27.98	3.189	15
131 Mi	29.52	3.026	21	29.57	3.021	10
131 Or	29.87	2.991	32	29.92	2.986	21
1-31	30.26	2.954	30	30.27	2.953	12
041	30.79	2.904	41	30.88	2.896	21
-132	32.39	2.764	19			

* I = relative intensity, calculated from X-ray patterns by assingning 100% intensity to the highest peak.

potassium feldspars. The presence of doublet reflections of potassium feldspars resulting from the 131 planes at 29.52° 2θ and the 1-31 planes at 30.26° 2θ and only a single reflection from the above two mentioned planes(131 reflection at 29.80° 2θ) indicates the presence of both triclinic microcline and monoclinic orthoclase, respectively. These characteristic diffraction lines (131 of Orthoclase and 131 of microcline) have been used for semi-quantitative determination of microcline and orthoclase as described by Somasiri and Huang (1971). Perthitic intergrowths of albite in potassium feldspars can be identified in powder samples by (1) pronounced separation of the -201 reflection of the K-feldspar phase at about 21.08° 2θ and the -201 reflection of the albite phase at about 22.06° 2θ and (2) the strong 002 peak of low albite at about 27.96° 2θ (Smith, 1974; Wright and Stewart, 1968). The 002 peak due to the albite phase of the perthitic intergrowth present in the samples is illustrated in Figure 21 and listed in tables 6 and 7. Reflections from the albite phase are present at 22.11±0.08° 2θ for the -201 reflection, 27.98±0.06° 2θ for the 002 reflection (Wright and Stewart, 1968). The composition of the potassium rich phase of perthite present in the samples can be calculated from the 2θ angle of the -201 reflection since the value of this angle is dependent on the unit-cell dimension which in turn is affected by composition of the feldspar (Wright, 1968). The measured mean of the -201 reflection from the samples studied is 21.04±0.02° 2θ. On the basis of these measured 2θ values and the diagrams of Wright (1968) relating composition to -201 peak location, the K-feldspar content ranges from 84 to 90% while content of Na-feldspar ranges from 10 to 16% by weight. These results are in good agreement with the composition determined by bulk chemical analysis. The angular separation between the -201 reflection of the potassium feldspars and that of the perthitic albite phase is related to the degree of exsolution or phase separation. In a non-exsolved solid solution these peaks merge into one. The angular separation of -201 peak for the different size fractions of the samples ranges from 1.00 to 1.08° which indicates a high degree of exsolution between the two feldspar phases (Marfunin, 1966).

The framework structure of feldspars consists of kinked ribbons of SiO_4 and AlO_4 tetrahedra with cations like K, Na, Ca or Ba occupying "cavities" produced by the alignment of kinked ribbons of Al and Si (Barth, 1969; Smith, 1974). The basic unit cell contains four units of the feldspar formula. For example, the unit cell of orthoclase contains four potassium atoms, four aluminum atoms, twelve silicon atoms and thirty two oxygen atoms. The pattern of distribution of silicon and aluminum in feldspars produces triclinic (ordered) symmetry with aluminum and silicon arranged in an orderly fashion in the kinked ribbons or monoclinic (disordered) symmetry where aluminum and silicon are distributed in a more or less random fashion (Barth, 1969).

These order-disorder forms are not highly descrete forms, rather they represent a continuum. As pointed out by Laves (1952), the degree of order in potassium feldspars reflects the temperature of crystallization. He also suggested that potassium feldspars of rather complete disorder exist just above the inversion temperature (about 700°C). Below 700°C some ordering tends to develop, until about 600°C where a high degree of order may be reached as in the naturally occurring microcline (Laves, 1952). Order-disorder relationships in feldspars are thought to be of importance because of their effects on the surface chemistry of feldspars and consequently on decomposition and possible alteration products (De Vore, 1959). Data on the structural states of feldspars in soils is also useful in understanding the origin of soil parent materials (Van der Plas, 1966; Somasiri and Huang, 1971).

Structural states in potassium feldspars can be ascertained by X-ray diffraction analysis (Goldsmith and Laves, 1954a&b). The potassium feldspar polymorphs are identified by the presence and location of the 131 and 1-31 reflections between 2θ values of 29° and 31° and the 130 and -130 reflections between 2θ values of 23° and 25° (MacKenzie, 1954). To determine the structural states of feldspars, the obliquity value or the departure of the feldspar structure from disordered to a more ordered triclinic polymorph is determined. The obliquity value is calculated according to the equation of Goldsmith and Laves

(1954b) where:

$$\text{Obliquity} = 12.5\ (d131 - d1\text{-}31)$$

The value for orthoclase (a monoclinic, disordered potassium feldspars) is equal to zero while that of maximum microcline (a triclinic, ordered potassium feldspar) reaches a value of 1 Å. Obliquity values of 0.2 to 0.5 Å are not accompanied by splitting of the 131 peak but rather broadening of the 131 towards the [illegible] peak. If the obliquity value is less than 0.2, this broadening can hardly be detected (Murfinin, 1966). Calculated obliquity values of the microcline polymorph in the selected samples used for illustrative purposes (Tables 6 and 7) range from 0.86 Å to 0.98 Å for the fine sand fraction and from 0.85 Å to 0.9 Å for the coarse and medium silt fraction. Values for various soil horizons and their associated size fractions are to be presented and discussed latter.

The specific gravity range of 2.59 to 2.72 g/cc separates albite, quartz and other plagioclases from heavy minerals and K-feldspars (Van der Plas, 1966). X-ray diffractograms for the 2.59-2.72 g/cc separates of both fine sand and coarse and medium silt are illustrated in Figure 22. Tables 8 and 9 are a tabulation of the X-ray data obtained for the two size fractions. X-ray data indicate the presence of large amounts of quartz in both fine sand and coarse and medium silt fractions as evidenced by the intense quartz peaks (100q and 101q). The diffraction patterns

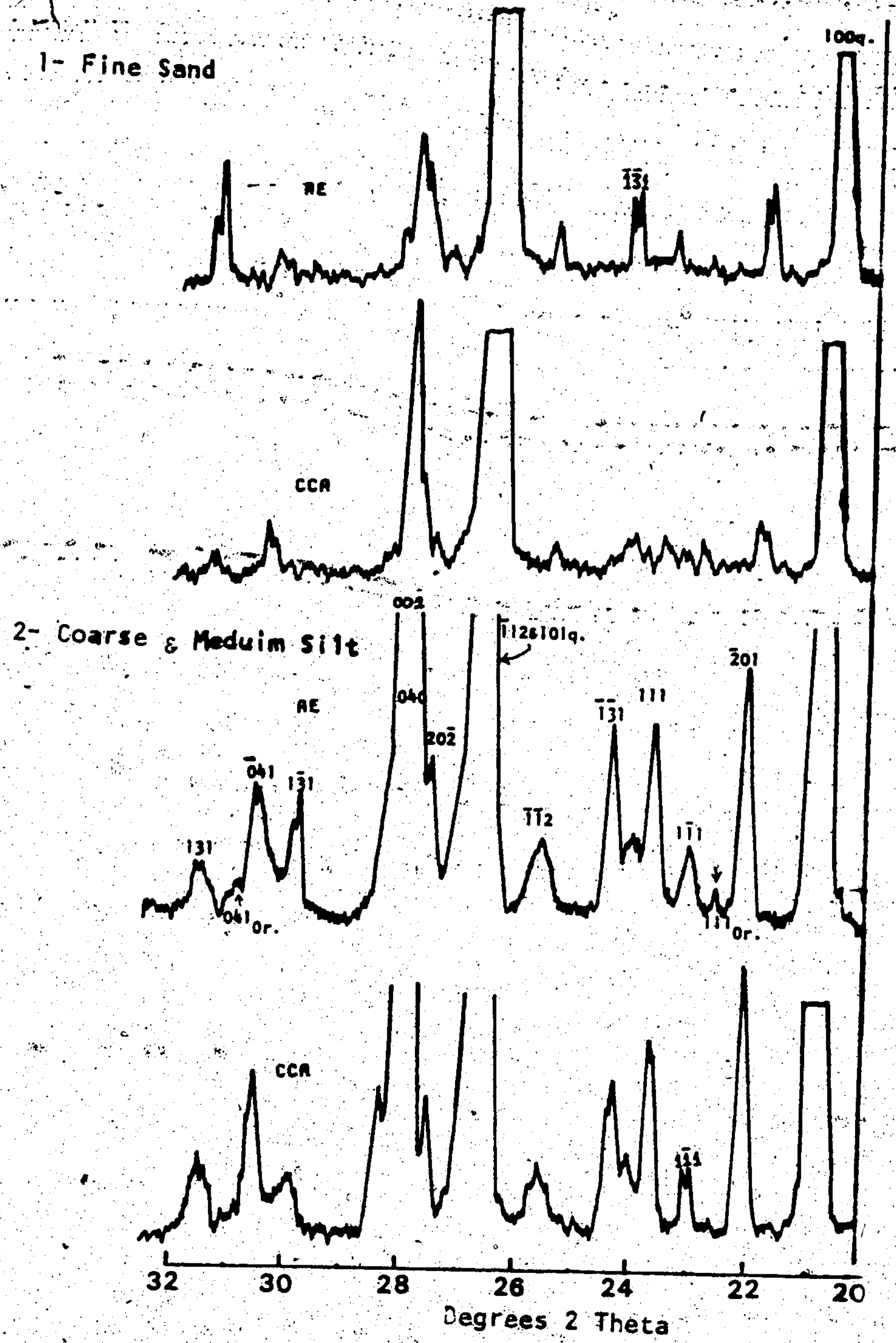


Figure 22. X-ray diffractograms for the 2.59-2.72 g/cc specific gravity separate from the Ae and Cca horizons of an Orthic Gray Luvisol soil (site 1) (all indices except as noted are for albite)

Table 8. Bulk chemical composition and X-ray diffraction characteristics of the 2.59-2.72 g/cc spcific gravity separate from the fine sand fraction of the Ae and Cca horizons of an Orthic Gray Luvisol (site 1).

Horizon	Ae	Cca
Bulk Composition	Ab67 Or13 An20 Ab77 An23*	Ab65 Or14 An21 Ab76 An24*

X-ray diffraction peaks

Miller Index	°2θ CuKα	d=Å	I**	°2θ CuKα	d=Å	I**
100 quartz	20.86	4.258	81	20.87	4.256	81
-201	21.98	4.044	14	21.98	4.044	9
1-11				22.98	3.870	8
111	23.68	3.757	8	23.58	3.773	8
-1-31	24.38	3.651		24.28	3.666	8
101 quartz &-112	26.63	3.347	100	26.63	3.347	100
-202	27.38	3.257	2			
040, 002	27.98	3.189	17	27.93	3.194	28
1-31	29.88	2.990				
0-41	30.48	2.933	7	30.48	2.933	8
131	31.48	2.842	15	31.48	2.842	6

* after correction for K-feldspar contamination.

** I=relative intensity, calculated from X-ray patterns by assigning 100% intensity to the highest peak

Table 9. Bulk chemical composition and X-ray diffraction characteristics of the 2.59- 2.72 g/cc specific gravity separate from the coarse and medium silt fraction of the Ae and Cca horizons of an Orthic Gray Luvisol (site 1).

Horizon	Ae	Cca
Bulk Composition	Ab63 Or21 An16 Ab80 An20*	Ab68 Or16 An16 Ab81 An19*

X-ray diffraction peaks

Miller Index	°2θ CuK_α	d=Å	I**	°2θ CuK_α	d=Å	I**
100 quartz	20.86	4.258	81	20.87	4.256	81
-201	22.01	4.038	27	22.06	4.029	28
-111 K-feld.	22.52	3.948	6	22.47	3.957	4
1-11	22.97	3.872	11	22.97	3.872	11
111	23.57	3.774	22	23.65	3.762	21
130 K-feld.	23.92	3.720	11	23.97	3.712	12
-1-31	24.27	3.667	22	24.27	3.667	18
-1-12	25.52	3.490	11	25.57	3.484	11
101 quartz &-112	26.63	3.347	100	26.62	3.349	100
-202	27.52	3.241	19			
002, 040	27.92	3.196	81	27.92	3.196	77
1-31	29.77	3.001	15	29.82	2.996	9
-041	30.47	2.934	16	30.47	2.934	18
041 K-feld.	30.92	2.892	7			
131	31.47	2.843	9	31.42	2.847	12

* after correction for K-feldspar contamination.

** I=relative intensity, calculated from X-ray patterns by assigning 100% intensity to the highest peak (sharp -201 and 040 reflections) also indicate the presence of albite. Occurrence of broad peaks (near 22.9° 2θ, 25.3° 2θ) or even partially resolved peaks (1-11 and -1-31) indicates some admixture of oligoclase and/or other more calcic feldspars. The diffractograms also indicate some contamination of potassium feldspars as evidenced by the broad peaks near 22.5°, 25.7° and 30.92° 2θ. The unexpected high content of KAlSi3O8 (Or.%) as indicated by the bulk chemical analysis of this separate (Tables 8 and 9) is probably due to contamination by perthitic potassium feldspars as well as possible contamination with mica. X-ray diffraction analysis of the silt separates and petrographic examination of the fine sand separates indicated that perthitic potassium feldspars falling in the 2.59-2.72 g/cc range, rather than mica, is the main reason for the unexpected high content of KAlSi3O8.

Though no detailed petrographic examination was conducted on the 2.59 to 2.72 g/cc separate, the few observations made of thin sections indicate that within this density separate albite, oligoclase and some more calcic feldspar are present. Some feldspar grains are perthitic in nature with the albite domains appearing larger than those of K-feldspar. The presence of albite, oligoclase and more calcic feldspars was also suggested by observations of twining and extinction angle measurements made on grain mounts of the total sand fraction from all horizon samples.

X-ray diffraction analysis can be used to differentiate low and high temperature plagioclase phases or to determine their chemical composition. The systematic change in lattice parameters with increasing calcium content allows for the determination of chemical composition by X-ray diffraction. In order to determine the composition of the plagioclase phase present using X-ray data, it is necessary to know whether the plagioclase is of low or high temperature origin. High temperature plagioclases are less frequently encountered in sediments and soils (Van der Plas, 1966; Huang, 1977) so it is possible to assume that the plagioclase phases present are of low temperature origin. This assumption could be true for feldspars which originated from granitic rocks from the Shield. The origin of feldspars in Cretaceous shales and sandstones is difficult to ascertain. The sedimentary rocks which underlie the plains region are marine or shallow water or terrestrial deposits. Sediments forming these deposits originated from normal weathering of sedimentary rocks and are brought into marine basins as a result of the rising Rocky Mountains to the west (Morton, 1964). Due to volcanic activity in the source area a proportion of the sediments was of volcanic origin, mainly volcanic ash which altered to bentonite as beds in shales. However, no zoning of feldspars was observed in the sand fractions of the soils nor in the very fine sand fraction separated from the one shale sample from the study area which suggests a non-volcanic origin (hence, low temperature form) for feldspars in till. Other problems which complicate X-ray characterization of the plagioclase fraction are the presence of more than one plagioclase phase and the presence of potassium oxide (Smith, 1956). X-ray diffraction data was used to determine the chemical composition of the albite-anorthite phase(s) assuming plagioclases were of low temperature form and that mixtures behave as a single phase. Measurements of degrees 2θ values of several reflections were made including reflections for the 1-11 at 111 at 23.6 ± 0.1°, 1-31 at 29.9 ± 0.25° 2θ, 131 at 31.4 ± 0.15° 2θ, and -132 at 33.85 ± 0.1° 2θ. The value of °2θ (131) - °2θ (1-31) ranges from 1.50 to 1.64° in the silt fraction and 1.63 to 1.67° in the sand fraction, which corresponds to 22 to 31% An and 30 to 33% An respectively (Smith, 1956). On the other hand, the value of °2θ (111) - °2θ (1-11) ranges from 0.64 to 0.68° in the silt fraction and 0.59 to 0.69° in the sand fraction which corresponds to 23 to 30% An and 18 to 32% An respectively (Smith, 1956). Finally, the value of ° 2θ (-132) -° 2θ (131) is 2.44 ± 0.01° in both the sand and silt fractions which corresponds to 29 to 31% An content for the plagioclase (Smith, 1956). In summary, these results suggest that the relative anorthite content of the soil plagioclase mixture ranges from 25 to 31% An. in the sand fraction while it ranges from 26 to 32% An. in the silt fraction which agrees, in general, with the relative anorthite content (on an orthoclase free basis) determined by bulk chemical analysis (Table 13 and 14, Appendix A). Such an agreement suggests that X-ray methods may be very useful in determining the composition of soil plagioclase.

To this point discussion was mainly centered on the application of X-ray diffraction data together with petrographic and chemical analysis as a means of characterizing the nature (composition, polymorphic forms, exsolution) of feldspars in soil separates. In the forthcoming section, feldspar mineralogy of the fine sand fraction from soils will be discussed in detail.

X-ray diffractograms of the 2.53 - 2.59 g/cc separates of the fine sand fraction (Figure 23 and Figures 5 to 9, Appendix A) indicate remarkable similarities between and within profiles. One noticeable difference within profiles is the increase with depth in resolution of the -201 reflection from the albite phase of perthitic K-feldspar. This trend was also accompanied by a substantial increase in th 002 reflection of the albite phase (Figure 23). Bulk chemical analysis shows that the lowest content of NaAlSi3O8 is always in the surface horizon as compared to the underlying horizons (Table 13, Appendix A) indicating that weathering of the albitic phase in the K-feldspar occurred to the greatest extent in the surface horizons of the profiles. This is in agreement with the Goldich stability series (1938) in which albite is less stable than potassium feldspars. Selective alteration of albite lamellae in perthite crystals of potassium feldspars were noticed to

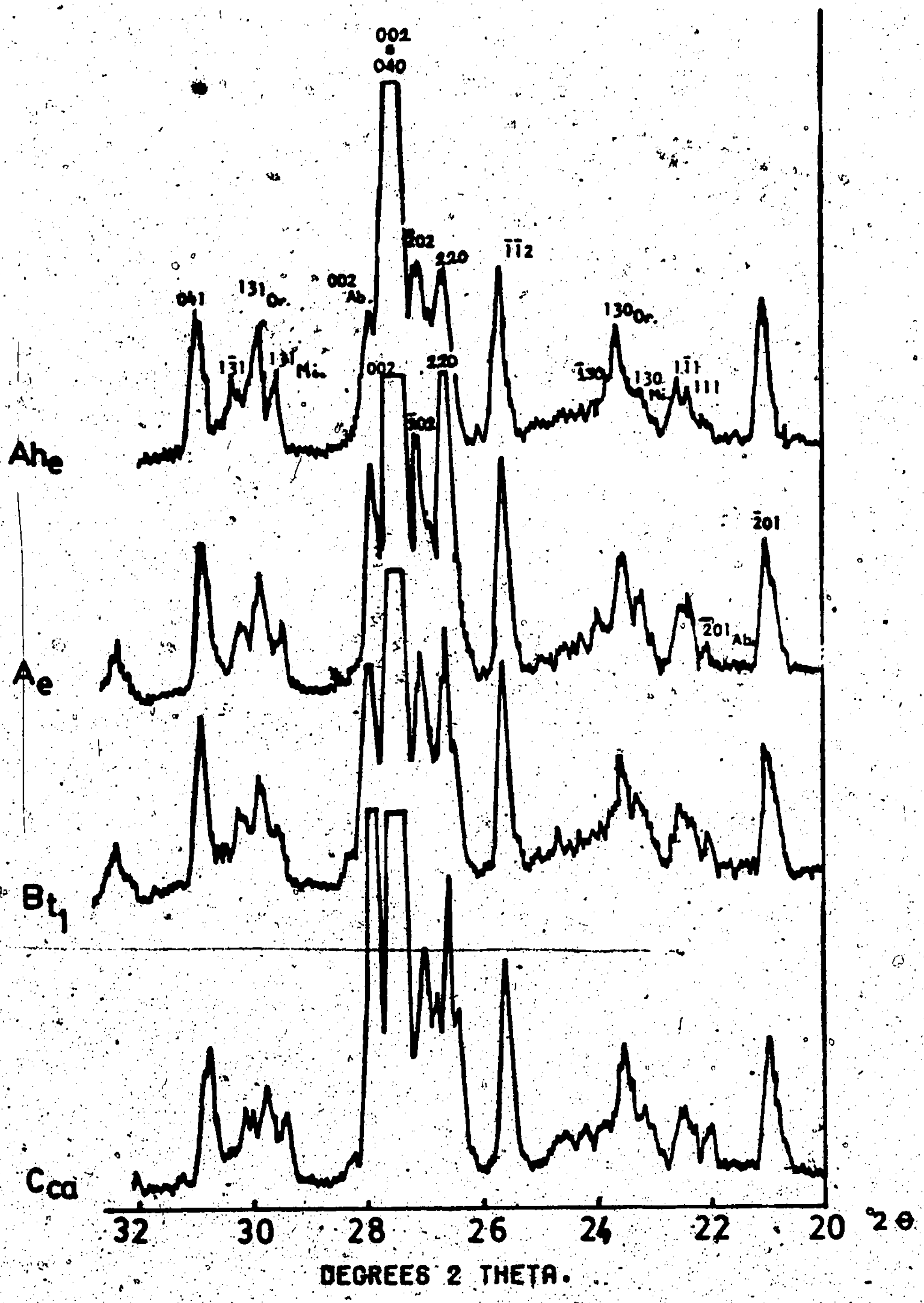


Figure 23. X-ray diffraction patterns for the 2.53-2.59 g/cc specific gravity separate from the fine sand fraction of the major genetic horizons of a Dark Gray Luvisol soil (site 3) (all indices except as noted are for K-feldspars).

occur in granites of South-West England (Exley, 1964).

Relative proportions of orthoclase to microcline polymorphs of K-feldspars in the sand fraction were determined from X-ray diffraction using the peak height ratios of the 131 reflection of orthoclase to the 131 reflection of microcline and comparing sample ratios with those of standard mineral mixtures (Figures 24 and 25) as described by Somasiri and Huang (1971). The results, presented in Table 10 along with the calculated obliquity values of microcline indicate that orthoclase constitutes 32 ± 6% of the sand-sized potassium feldspars on the average while microcline makes up the remainder. Variations within profiles are about up to 15% of the means. The exceptions are the Orthic Gray Luvisol profile at site 4 and the Dark Gray Luvisol at site 2 where the cause of the large variations is due primarily to unusually high or low contents for only one horizon sample from each. Original variations in both content and size of orthoclase and microcline in the till since deposition may explain the results. The erratic behaviour of the relative quantity of orthoclase to microcline with depth suggests pedogonesis has not altered the relative quantities of orthoclase and microcline. The obliquity values of the microcline from the fine sand fraction from the major genetic horizons range from 0.75 to 0.98 Å with an average of 0.86 ± 0.06 Å. A range of obliquity values between 0.82 and 0.87 Å was reported for K-feldspar separates from tills in Saskatchewan

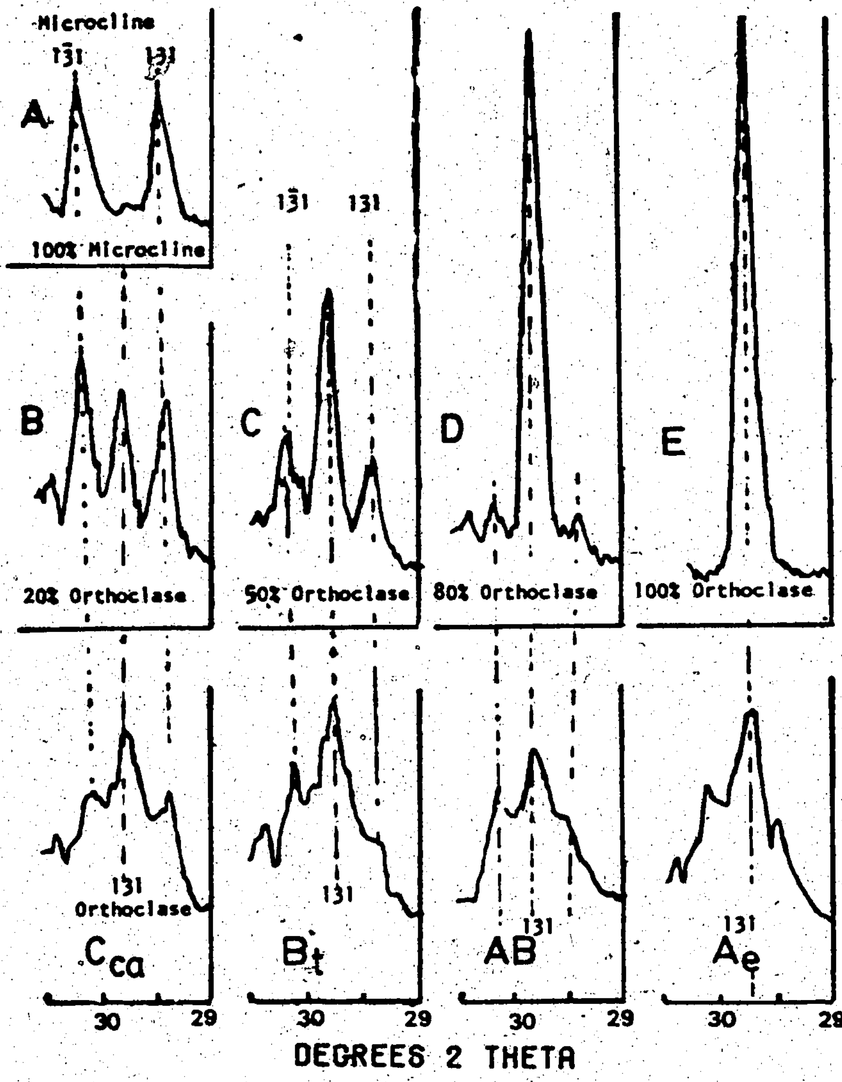


Figure 24. X-ray diffractograms showing variations in 131 and $1\bar{3}1$ of microcline versus 131 of Orthoclase for standard mineral mixtures (A–E); and variations occurring within horizons of an Orthic Gray Luvisol (site 1).

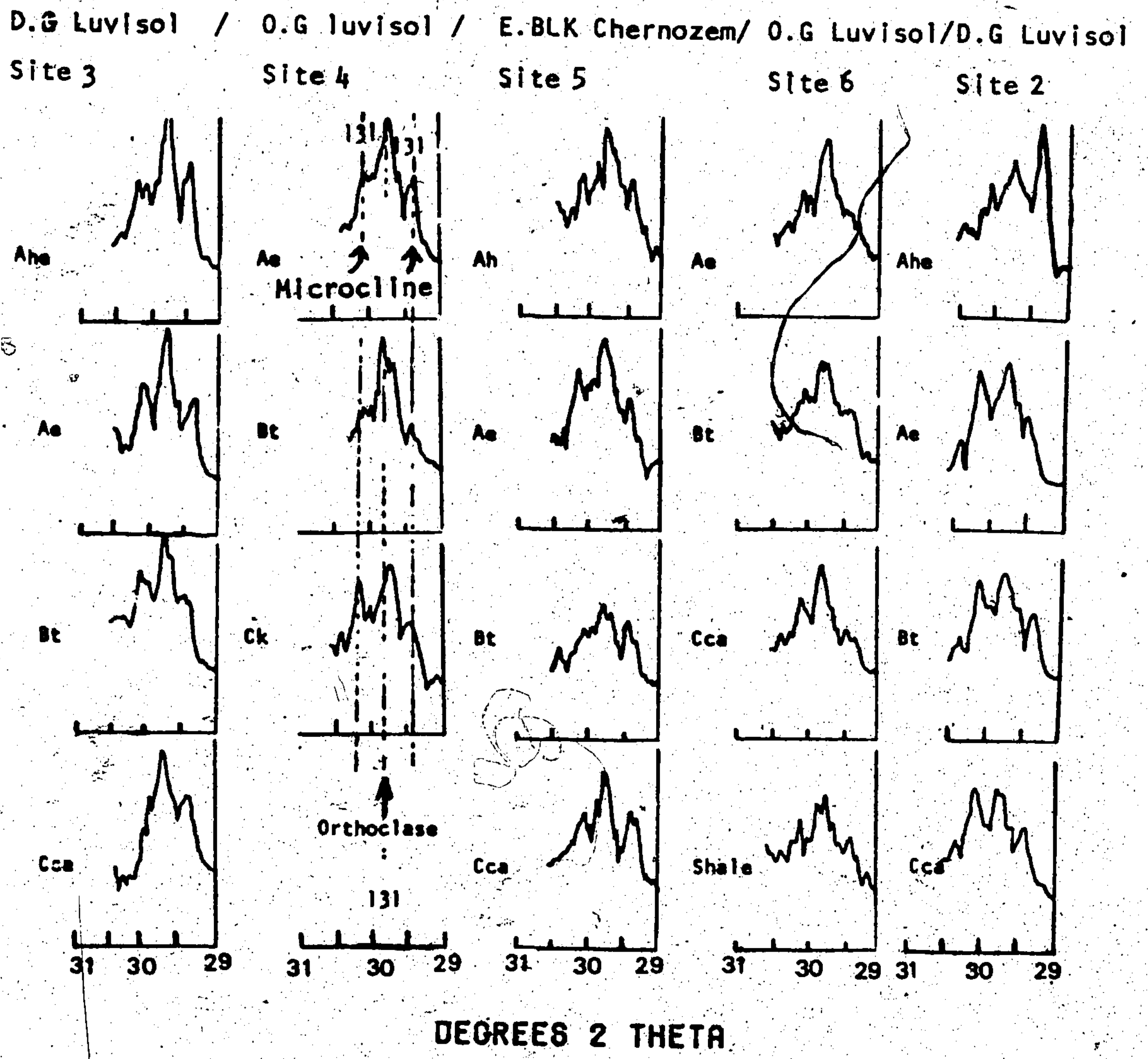


Figure 25. X-ray diffractograms showing variations in 131 and $\bar{1}31$ of microcline versus 131 of orthoclase in the sand fractions from the major horizons within sites 2, 3, 4, 5 and 6.

Table 10. Relative proportions of orthoclase and microcline in K-feldspars and the obliquity value of microcline in the fine sand fraction separated from the major genetic horizons,

Soil	Horizon	Orthoclase		Microcline		Obiquity value	
		%	Mean ± S.Dev	%	Mean ± S.Dev		Mean ±S.Dev.
Orthic Gray Luvisol (site 1)	Ae	36	34 ± 5	64	66 ± 5	0.86	0.88 ± 0.03
	AB	32		68		0.86	
	Bt2	40		60		0.86	
	Cca	28		72		0.92	
Dark Gray Luvisol (site 2)	Ahe	13	27 ± 9	87	73 ± 9	0.84	0.87 ± 0.04
	Ae	33		67		0.86	
	Bt2	30		70		0.93	
	Cca	30		70		0.84	
Dark Gray Luvisol (site 3)	Ahe	30	32 ± 2	70	68 ± 2	0.86	0.87 ± 0.05
	Ae	34		66		0.93	
	Bt1	32		68		0.89	
	Cca	30		70		0.81	
Orthic Gray Luvisol (site 2)	Ae	30	37 ± 7	70	63 ± 7	0.86	0.80 ± 0.09
	Bt1	44		56		0.85	
	Ck	36		64		0.70	
Eluviated Black Chernozem (site 5)	Ah	30	29 ± 4	70	71 ± 4	0.88	0.82 ± 0.06
	Ae	33		67		0.85	
	Bt1	23		77		0.75	
	Cca	28		72		0.81	
Orthic Gray Luvisol (site 6)	Ae	39	36 ± 5	61	64 ± 5	0.85	0.88 ± 0.09
	Bt	31		69		0.81	
	Cca	39		61		0.98	
Shale		28		72		0.92	
Grand mean ± S.Dev			32 ± 6		68 ± 6		0.86 ± 0.06

(Somasiri and Huang, 1971) and said to approximate values obtained for K-feldspars of Shield rocks. The average of obliquity values for this study lies within the range reported for K-feldspars of Saskatchewan and suggests similarities in the origin of K-feldspars of Saskatchewan tills and those of this study area.

Scanning electron microscope examination of selected samples from the 2.53 to 2.59 g/cc separates show morphological features similar to those of feldspars in the <2.53 g/cc separate of the fine sand fraction (Plate 6). Extensive solution pitting and formation of secondary minerals such as kaolinite are the main features observed (Plate 6).

In representitive X-ray diffractograms of the 2.59 to 2.72 g/cc separates (Figure 26 and Figure 10, Appendix A) dominance of quartz is suggested by the presence of intense 100 and 101 reflections at 20.85 and 26.65° 2θ (4.26 Å and 3.35 Å), respectively. The presence of relatively small amounts of plagioclase is suggested by small peaks which mainly resemble d-spacings of low albite. As already shown (Figure 20) the admixture of some other plagioclases such as oligoclase and/or more calcic feldspars is difficult to ascertain but the broad peaks and splitting of the 1-11 and -1-13 peaks observed in the original diffraction patterns suggests other plagioclases are present. Presence of the 111 and 041 reflections of K-feldspars at about 22.5 and 30.8

Description of Plate 6

A and B: Weathered feldspar grain displaying striated perforations. The solution pitting as seen in micrograph B could represent differential weathering along twin planes or selective weathering of one phase of a perthitic intergrowth.

C to E: Formation of secondary mineral material on the surface of feldspars as evident by numerous stacks of irregular shaped particles (C and D) some of which display a weak hexagonal shape. Differential surface weathering is displayed by the feldspar in micrograph E.

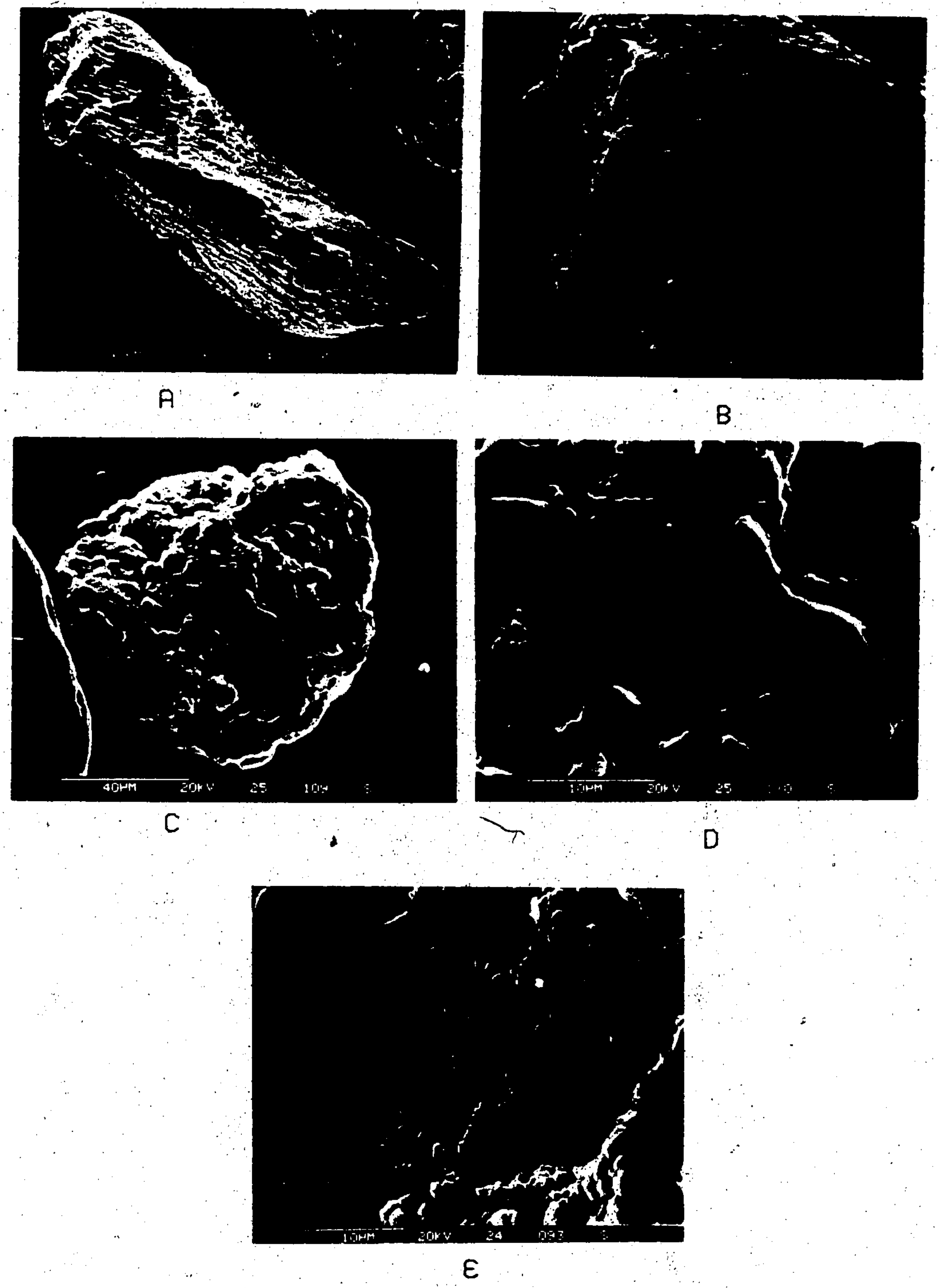


Plate 6. Scanning electron micrographs of sand-sized alkali feldspar grains separated from the Ae horizon (A, B and E) and the Cca horizon (C and D) of an Orthic Gray Luvisol (site 6).

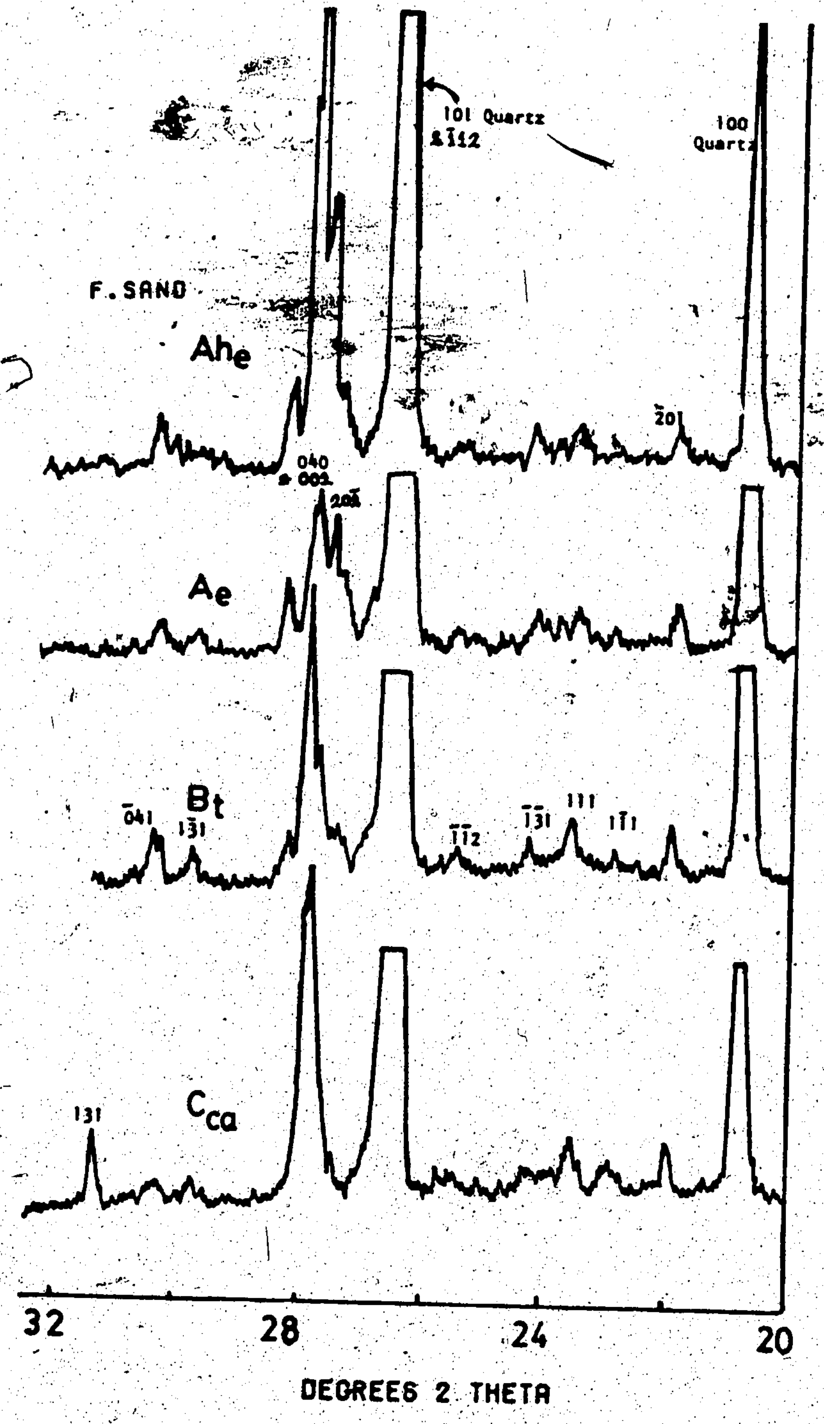


Figure 26. X-ray diffraction patterns of the 2.59-2.72 g/cc specific gravity separate from the fine sand fraction of the major genetic horizons of a Dark Gray Luvisol (site 3) (all indices except as noted are for albite)..............

2θ, respectively on the original XRD patterns suggests that K-feldspars and/or oligoclase are probably present in most of the samples, especially in the surface horizons of all the profiles. The bulk chemical analysis (Table 14, Appendix A) indicates the greatest amount of $KAlSi_3O_8$ is present in the same horizons where X-ray diffraction patterns display peaks at about 22.5 and 30.8° 2θ. These results along with other evidence suggested earlier (petrographic examination) indicate that the 2.59 to 2.72 g/cc separate is contaminated with K-feldspars and substantiate the validity of recalculating the mole per cent of the separate so that plagioclase composition can be expressed on a $KAlSi_3O_8$ free basis(Table 14, Appendix A). This recalculation is in order since rock-forming plagioclases rarely contain more than 5% $KAlSi_3O_8$ (Muir, 1962). Furthermore, antiperthitic plagioclases (plagioclases with K-feldspar inclusions) occur in relatively high temperature rocks (granulite facies, volcanic and contact metamorphic) (Sen, 1959). These rock types are not likely contributors to the tills of this study.

Recalculated values indicate that feldspars in the 2.59 to 2.72 g/cc separate contain an average of 78.8 ± 2.6% $NaAlSi_3O_8$ and 22.2 ± 2.6% $CaAl_2Si_2O_8$. The variations between profiles are very small (3% of the means) which indicate remarkable similarity in the composition of the plagioclase fractions from all profiles. The composition of the K-feldspar fraction (2.53 to 2.59 g/cc separate) is also similar among the six profiles (Table 14, Appendix A).

Scanning electron microscopic examination of the surface texture of quartz and plagioclase feldspar was conducted on selected samples from the 2.59 to 2.72 g/cc separates. Solution pitting was an extensive surface feature (Plate 7) as in the case of K-feldspars. Weathering products are evident on the surface of most plagioclase grains. The susceptibility of plagioclase to chemical alteration and mechanical abrasion is evident from rounding of the grain edges and their spongy surface appearance (Plate 7-A and B). The spongy, pitted and rounded appearance as displayed in A and B is typical of the appearance of most of the sand-sized plagioclases. In contrast, K-feldspars usually display cleavage faces, sharp edges and less extensive pitting.

Surface texture of quartz grains is claimed to be diagnostic or reflective of the environmental history and origin of different geological deposits (Margolis, 1968; Krinsley and Donahue, 1968; Coach and Krinsley, 1971; Marzolf, 1976). Surface morphological features of quartz grains associated with the glacial environment as described by Krinsley and Donahue (1968), Coach and Krinsley (1971) and Krinsley and Doornkamp (1973) were observed in many of the sand grains examined in this study (Plate 8). These features include large variations in size of conchoidal breakage patterns, arc-shaped steps and imbricated breakage blocks. These features are probably related to the grinding

Description of Plate 7

A: Weathered plagioclase grain showing some solution pitting. Many of the dissolution pits are tiny unlike the large cavities in K-feldspars. Cleavage in the grain is obscured by the roundness of the edges probably due to chemical weathering during and after glacial transport.

B and C: Extensive surface dissolution without pitting and secondary mineral formation on a plagioclase feldspar grain. The secondary material (micrograph C which is a close up of B) appear as stacks of elongated or platy crystals. The platy and somewhat edge-curled morphology of aggregates seen in the lower right of micrograph C strongly suggests phyllosilicate material.

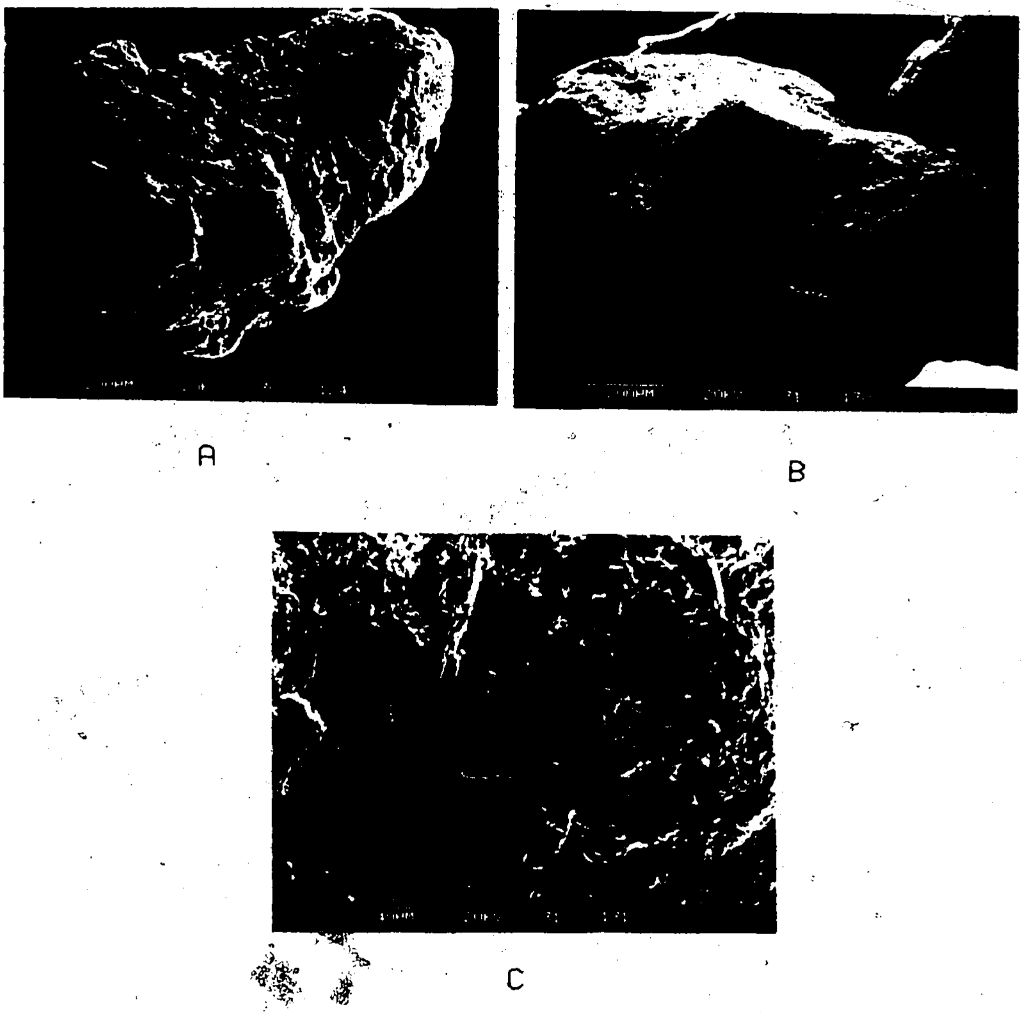
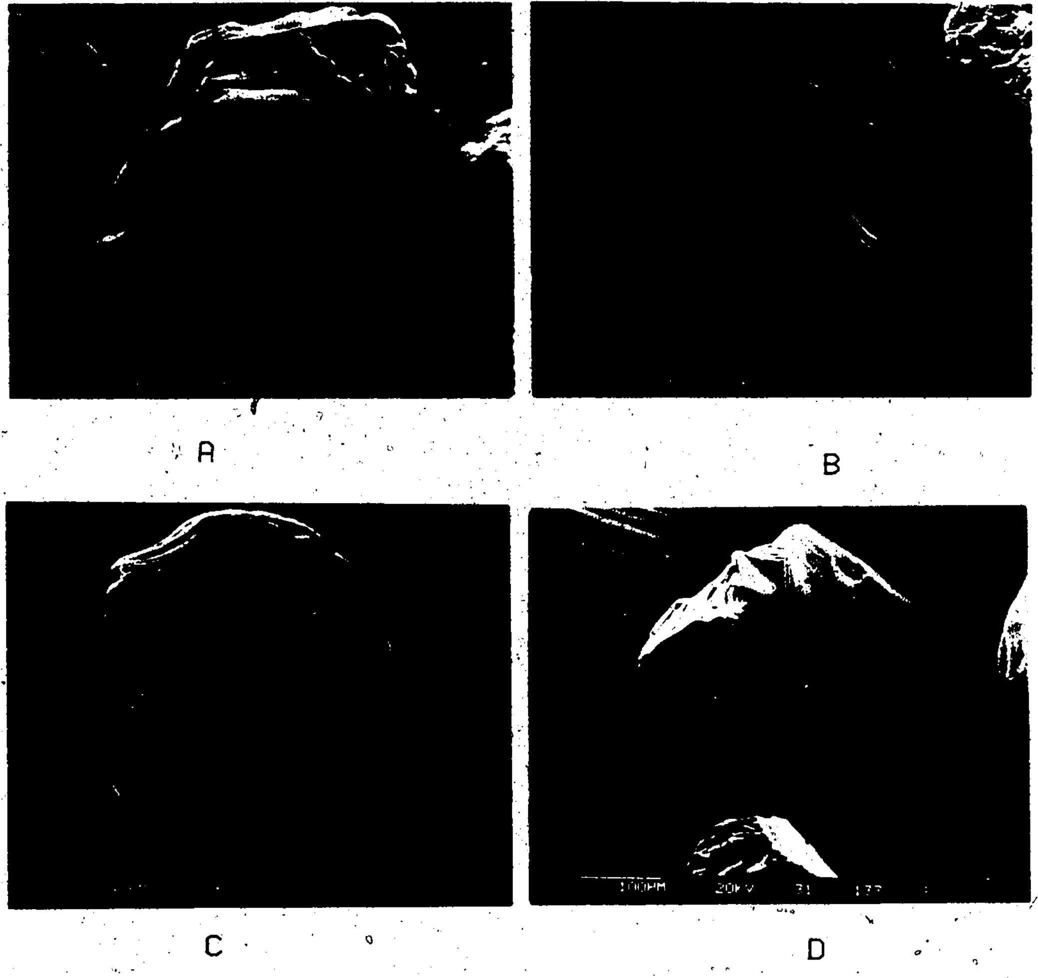


Plate 7. Typical scanning electron micrographs of plagioclase grains separated from the Ae (A) and AB (B and C) horizons of an Orthic Gray Luvisol (site 6).

Decription of Plate 8

A to D: Angular quartz grains displaying characteristic conchoidal fractures. Most edges are sharp and show little rounding and faces appear fresh. Conchoidal fracture patterns are different in shape and direction from extremely irregular to arc like steps.like those shown in the middle right corner of micrograph B and the upper left corner of micrograph D. The fresh appearance of fracture faces are indicative of minimal if any chemical dissolution or precipitation and are typical of grains that have been physically abraded by glacial grinding and transport.



A

B

C

D

Plate 8. Scanning electron micrographs of quartz grains separated from the Ae (A, B and C) and AB (D) horizons of an Orthic Gray Luvisol (site 6).

action grains are subjected to during glacial transport. Some quartz grains show a shell like appearance (Plate 8, C and D) resulting from several curved striations on fracture faces. These striations are generally comparable in appearance to the semi parallel steps observed by Krinsley and Donahue (1968) and are probably the cause of shear stress on the surface of the particles during glacial transport. The surface morphologies of all quartz grains examined where not compatible in appearance to quartz grains of an eolian origin described by Krinsley and Doornkamp (1977).

Petrographic examination of the coarse and medium and fine sand fractions confirm the results obtained by X-ray analysis concerning the nature of K-feldspars present in the 2.53 to 2.59 g/cc separate. Plate 9 illustrates the presence of different types of perthitic K-feldspars and microcline. Rock fragments containing several feldspar phases as well as other minerals(e.g., quartz) were also evident particularly in the coarse and medium sand fraction. Although most feldspar grains appear fresh and unaltered, some do display a weathered appearance as evidenced by the occurrence of alteration products(sericite) as seen in thin sections. Differential weathering of albite lamelle in perthetic K-feldspars was also observed (Plate 9).

The heavy minerals (specific gravity separate >2.72 g/cc) in the fine sand fraction consist of amphiboles

Description of Plate 9

A: Hair perthites.

B(left): Shows three different textures of albite intergrowths: replacement (upper right grain), rod like (center grain) and penetrating blebs (lower right grain). In the micrographs, albite intergrowths appear white while the K-feldspar matrix is gray. The upper right grain in micrograph B(right) represents albite veins (white) forming a braided structure of perthite. Film albite (narrow white strings) is present in the lower right large grain of this micrograph.

C(right): Veins of albite intergrowths (white) in partially altered K-feldspar (gray). The speckled appearance of the large grain (micrograph C, left) is indicative of feldspar alteration or irregular albite intergrowths.

D(left): A rock fragment of perthite (with albite as veins) and minor quartz (white spots). These rock fragments were abundant in the feldspar separate (2.53 to 2.59 g/cc) of the coarse and medium sand fraction. Micrograph D(right) represents a common vein perthite in which the albite lamellae of the perthitic K-feldspar has weathered. The irregular spots of high birefringence are muscovite.

E: Cross-hatched microcline grain (section cut parallel to 001) typical maximum microcline and common in the feldspar separate (2.53 to 2.59 g/cc).

F: Intergrowths of vermicular quartz (white) in K-feldspars (gray).

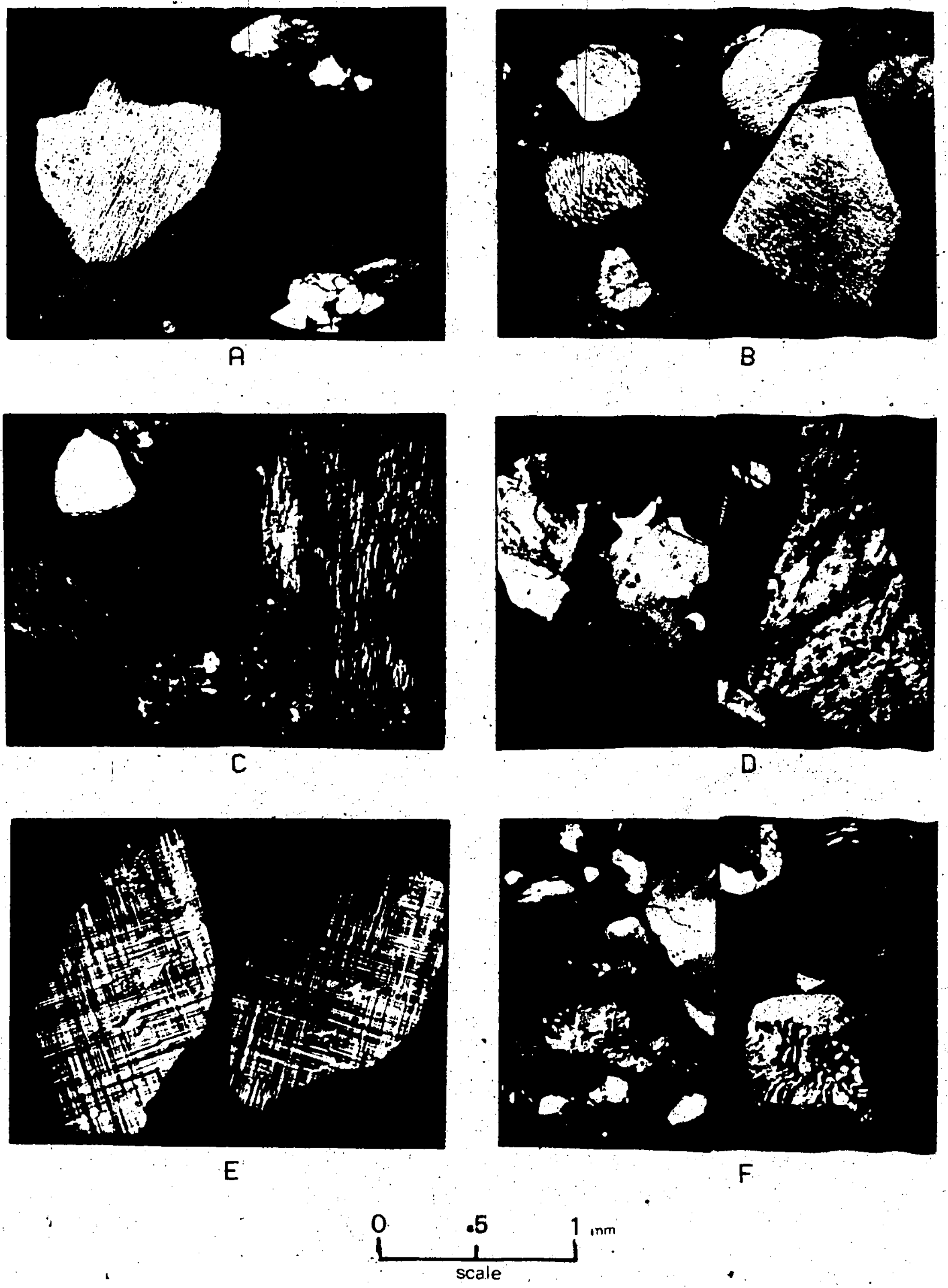


PLate 9. Typical perthitic K-feldspars,maximum microcline and intergrowths of quartz in K-feldspars as seen under the petrographic microscope (X-nicols).

(hornblende, glaucophane), iron oxides (magnetite, limonite and geothite), mica (mainly biotite), garnet, pyroxenes (augite) and small amounts of apatite, tourmaline, chlorite, epidote and zircon. This mineral assemblage is qualitatively similar to those observed for tills in Alberta (Bayrock, 1962; Twardy, 1969). The relative amounts of the dominant heavy minerals present in the major genetic horizons as estimated from thin section examinations are listed in Table 11. Total content of heavy minerals is lower in A horizons and generally lower in B horizons as compared to contents in C horizons and indicates weathering during soil development has affected heavy mineral distribution. Variations in amounts of individual minerals are not pronounced, however, detailed quantitative information was not obtained in this study. Generally, biotite appeared to be less abundant in the surface horizons of most profiles compared to amounts in C horizons and amphiboles appeared to be more abundant in C horizons as compared to A and B horizons. Furthermore, sand-sized aggregates of secondary iron oxides and quartz were present in relatively high amounts in the B horizons as compared to other horizons.

The overall composition of the major mineral species in fine sand fractions for the individual profiles are summarized in histograms (Figures 27 to 32 and Tables 7 to 12 Appendix - A). The values are calculated on a carbonate-free basis. Thus, the representation of carbonate in the graphs causes the totals to be in excess of 100% for

Table 11. Total content of heavy minerals in the fine sand fraction and their relative distribution within the >2.72 g/cc separate.

		relative amounts*					Heavy
Soil	Horizon	Amphiboles	Iron oxides	Biotite	Garnet	Pyroxenes	mineral content (%)
Orthic Gray Luvisol (site 1)							
Orthic Gray	A	5**	5	3	2	1	2.2**
Luvisol	Bt	5	4	4	2	1	2.2
(site 1)	Cca	5	4	5	2	1	2.7
Dark Gray Luvisol (site 2)							
Dark Gray	A	6	4	4	1	1	2.0
Luvisol	Bt	6	4	4	1	1	2.9
(Site 2)	Cca	6	4	4	1	1	2.3
Dark Gray Luvisol (site 3)							
Dark Gray	A	5	4	3	4	1	2.4
Luvisol	Bt	5	4	3	3	1	2.4
(site 3)	Cca	5	3	4	3	1	2.8
Orthic Gray Luvisol (site 4)							
Orthic Gray	A	4	6	2	3	1	1.6
Luvisol	Bt	4	6	2	3	1	1.7
(site 4)	Ck	4	6	2	3	1	2.6
Eluviated Black Chernozem (Site 5)							
Eluviated	A	5	4	3	3	1	2.2
Black	Bt	5	4	3	3	1	2.3
Chernozemic	Cca	5	4	3	3	1	2.3
(site 5)							
Orthic Gray Luvisol (site 6)							
Orthic Gray	A	5	4	2	4	1	1.2
Luvisol	Bt	5	5	2	4	1	1.4
(Site 6)	Cca	5	5	2	4	1	1.6

*Relative amounts: 1 = >5%; 2 = 5-10%; 3 = 10-20%; 4 = 20-30%; 5 = 30-45%; 6 = 45-60%

*average for all sub-horizons within each of the major horizons

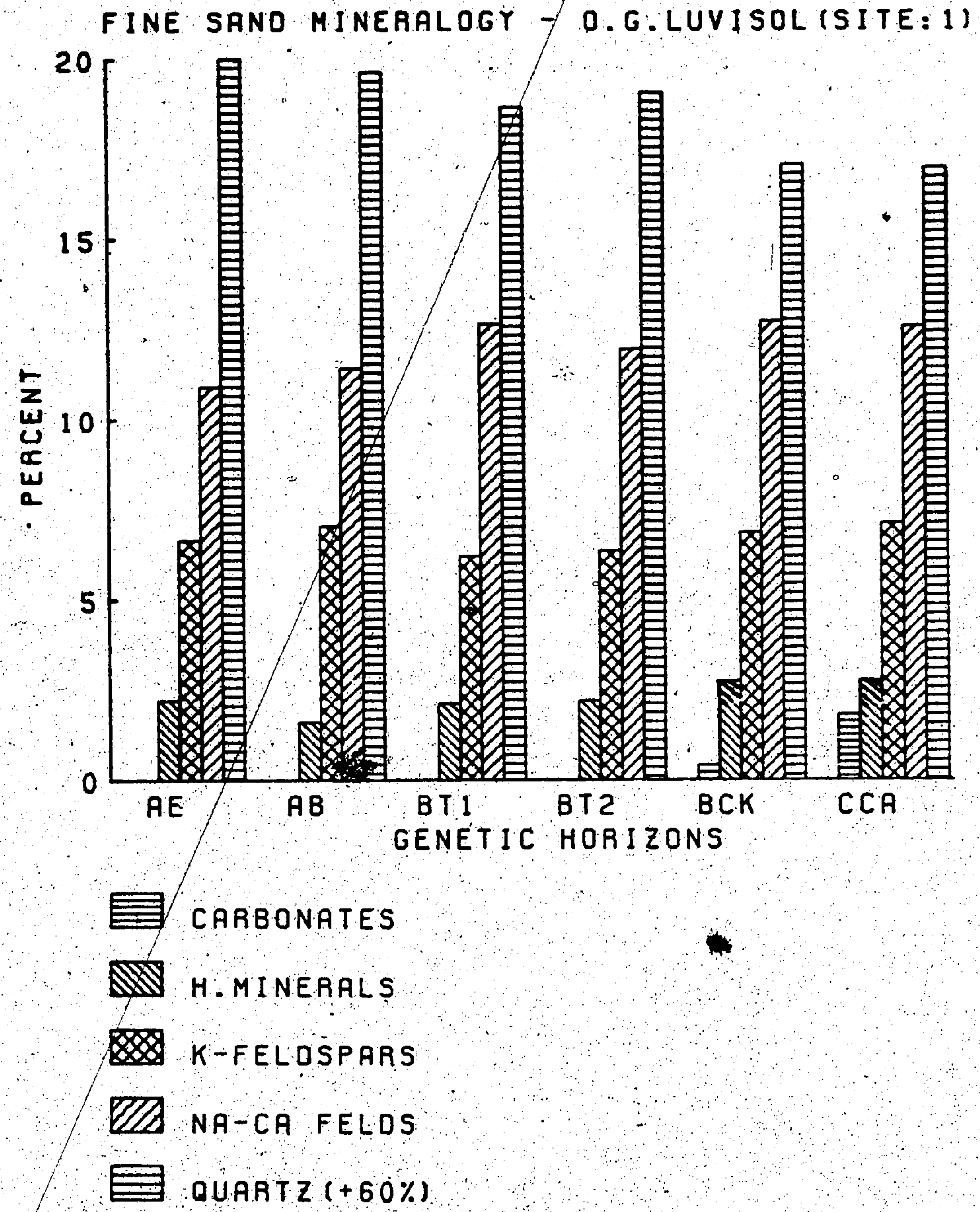


Figure 27. Histograms showing mineralogical constituents (amounts expressed in per cent on a weight basis) of the fine sand fraction separated from the various genetic horizons of an Orthic Gray Luvisol (site 1)............

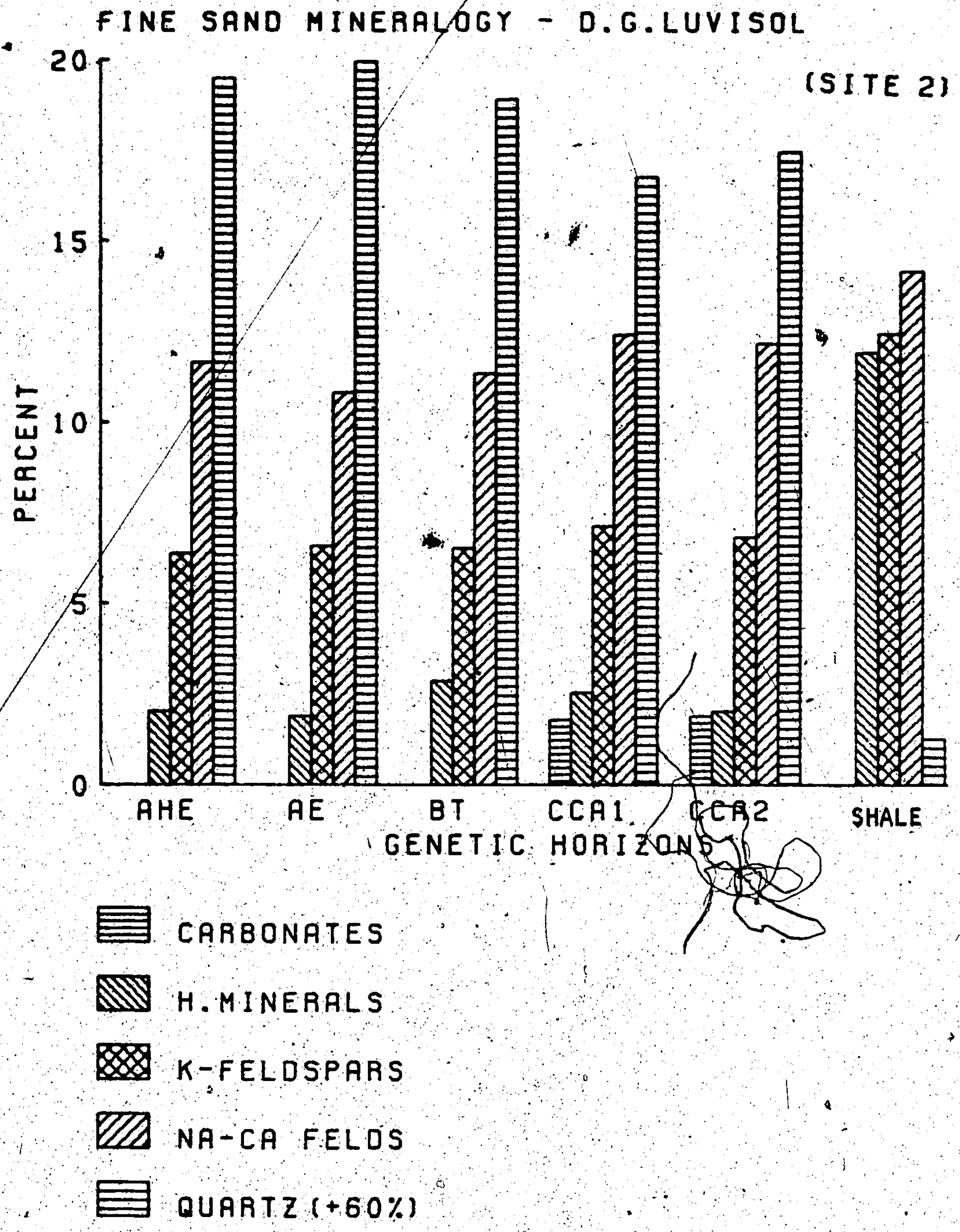


Figure 28. Histograms showing mineralogical constituents (amounts expressed in per cent on a weight basis) of the fine sand fraction separated from the various genetic horizons of a Dark Gray Luvisol (site 2)...............

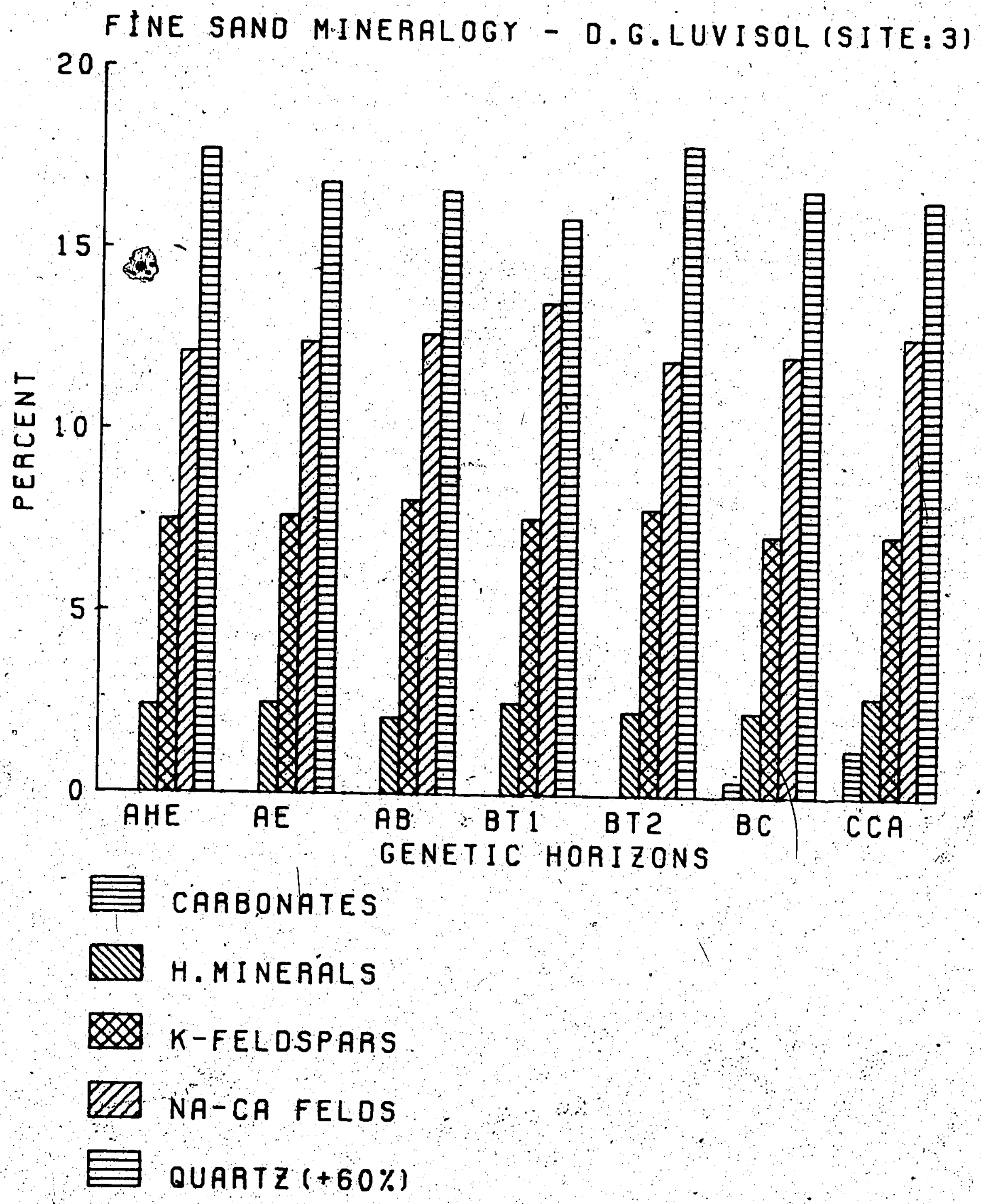


Figure 29. Histograms showing mineralogical constituents (amounts expressed in per cent on a weight basis) of the fine sand fraction separated from the various genetic horizons of a Dark Gray Luvisol (site 3)..............

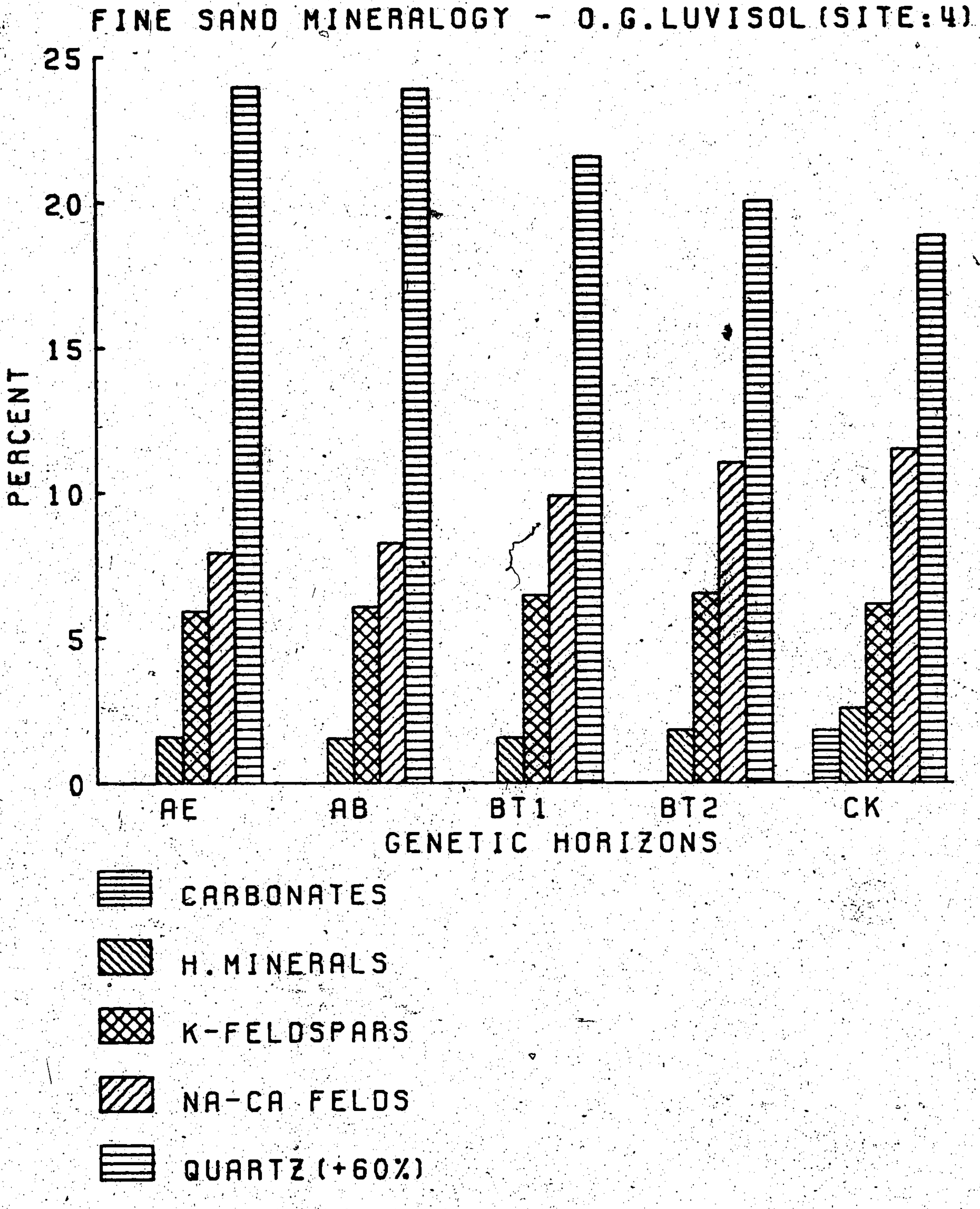


Figure 30. Histograms showing mineralogical constituents (amounts expressed in per cent on a weight basis) of the fine sand fraction separated from the various genetic horizons of an Orthic Gray Luvisol (site 4)............

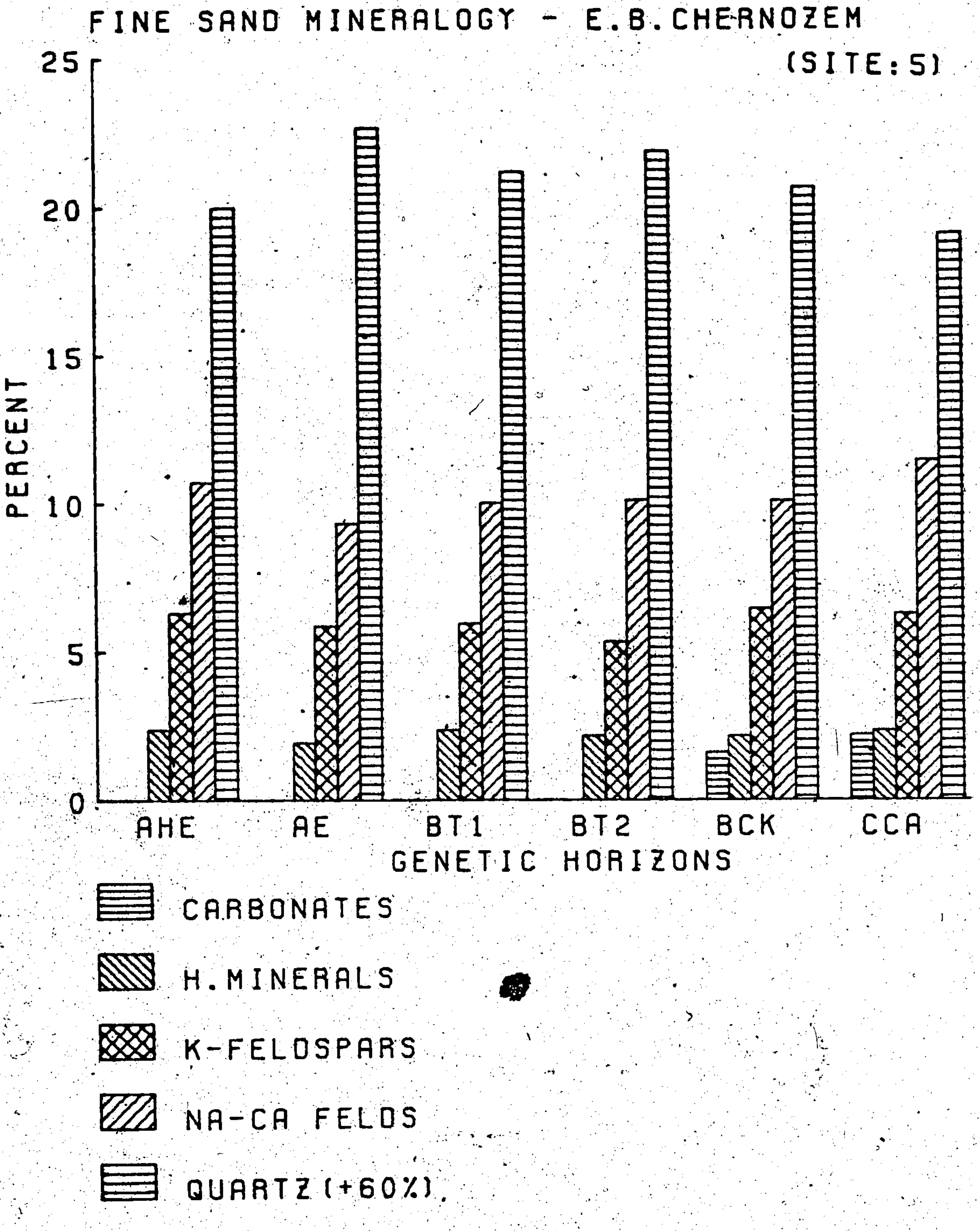


Figure 31. Histograms showing mineralogical constituents (amounts expressed in per cent on a weight basis) of the fine sand fraction separated from the various genetic horizons of an Eluviated Black Chernozem (site 5)......

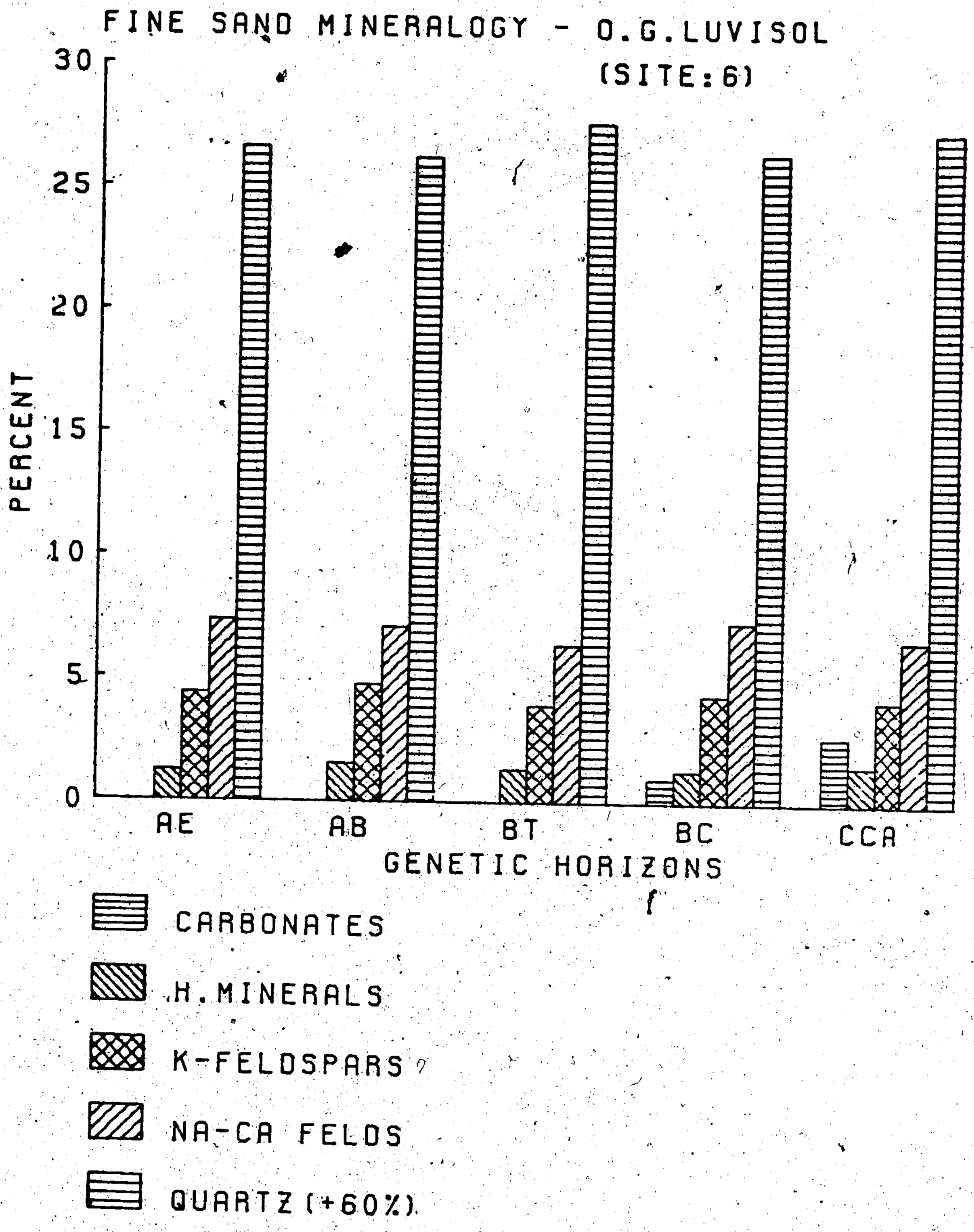


Figure 32. Histograms showing mineralogical constituents (amounts expressed in per cent on a weight basis) of the fine sand fraction separated from the various genetic horizons of an Orthic Gray Luvisol (site 6)............

carbonate containing horizons. Feldspar data presented in Figures 27 to 32 were obtained from the combined results of the 2.53 to 2.59 g/cc and 2.59 to 2.72 g/cc density separates. The graphs illustrate a marked degree of similarity in the bulk mineralogical composition of the fine sands from five pedons (sites 1,2,3,4, and 5). With increasing depth in the profiles, the heavy minerals and Na-Ca feldspars show a slight increase which is matched by a corresponding decrease in the content of quartz. Differences in potassium and Na-Ca feldspar contents among profiles are not marked except for the profile at site 6. The content of potassium feldspar at sites 1 to 5 ranges from 6 to 8% whereas site 6 has less than 5 percent. Content of sodium-calcium feldspars ranges from 10 to 14% in profiles at sites 1 to 5 and less than 7% at site 6. These variations may be attributed to possible differences in the source area of the glacial drift from the shield and/or mode of deposition. Feldspars present in the light mineral portion (specific gravity <2.7 g/cc) of the fine and very fine sands from tills in the same general area were found to vary from 6 to 8% for potassium feldspars and 9 to 15% for soda calcic feldspars (Pawluk and Bayrock, 1969). These amounts are in agreement with those found in this study.

c. Coarse and medium silt fraction

Characterization of the coarse and medium silt fraction was conducted using the same techniques utilized in the fine sand fraction. The fine silt was analyzed separately because of expected difficulties in heavy liquid separation resulting from slow sedimentation rate and/or aggregation (Jackson, 1975).

On the basis of X-ray diffraction analysis the light and heavy mineral separation appears to be effective in dividing the coarse and medium silt fraction into K-feldspars in the 2.53 - 2.59 g/cc separate (Figure 33), quartz and plagioclases in the 2.59 - 2.72 g/cc separate (Figure 34) and amphiboles, mica, talc and possibly chlorite and vermiculite in the >2.72 g/cc separate (Figure 35). The segregation of mica into the >2.72 g/cc separate was almost complete as indicated by the presence of a 10 Å peak in the >2.72 g/cc separate (Figure 35) but not in either the 2.53 - 2.59 g/cc or 2.59 - 2.72 g/cc separates (Figures 33 and 34).

In the >2.72 g/cc separate, the presence of amphiboles is verified by two main peaks at 8.45 Å and 3.15 Å while peaks at 10 Å and its higher order reflections indicate the presence of mica. The 14 Å peak could be attributed to both vermiculite and chlorite. Vermiculite could not be confirmed without K-saturation and hydration treatments. The persistance of the 14 Å peak after heating indicates the presence of chlorite. The presence of kaolinite was ruled out because of the absence of the 002 reflection of kaolinite at 3.58 Å. The relative proportions of these minerals in the >2.72 g/cc separate (calculated from intensities of their characteristic peaks using the method

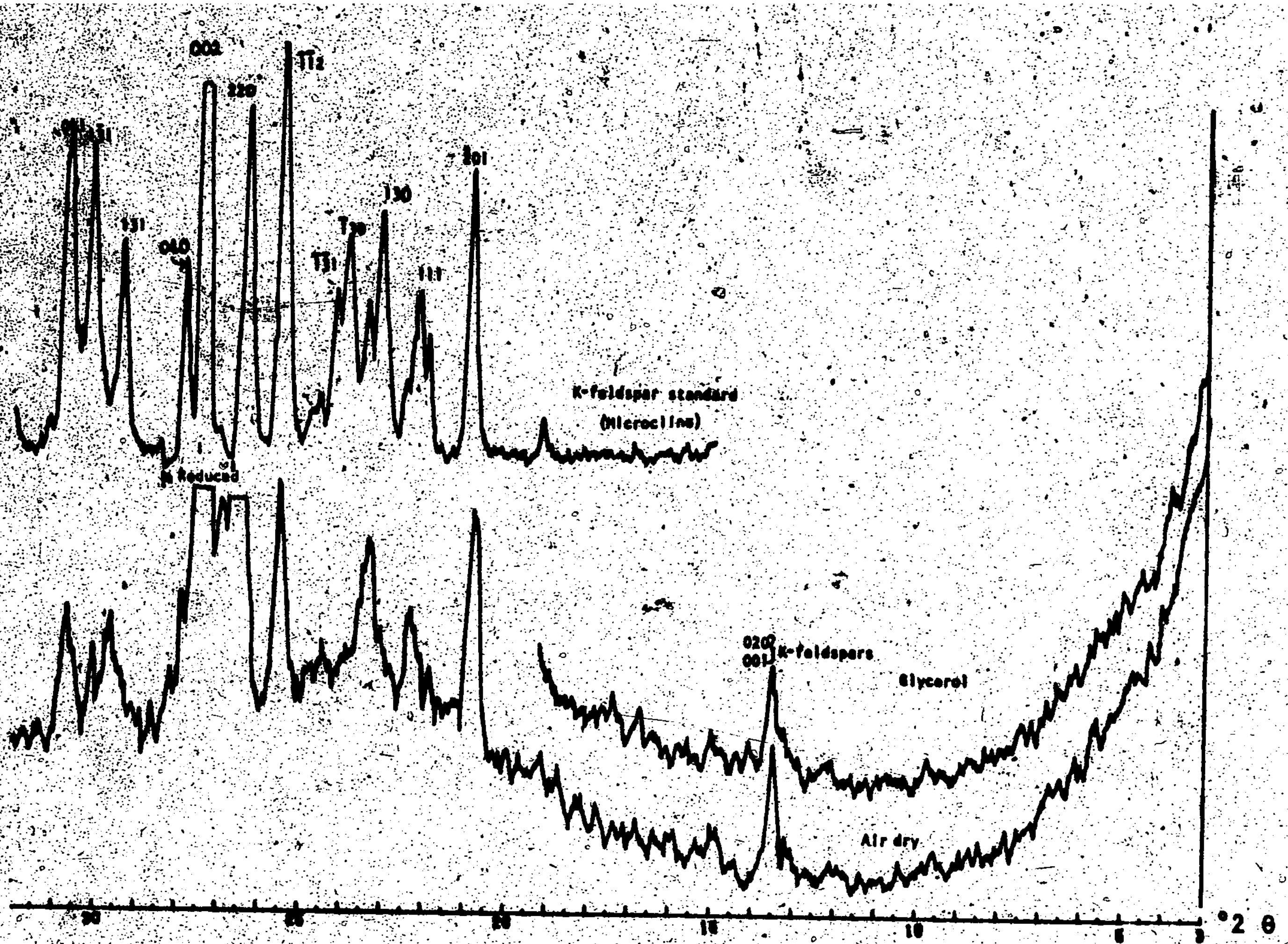


Figure 33. Typical X-ray diffractograms for the 2.53 to 2.59 g/cc separates from the coarse and medium silt fraction. The upper diffractogram represents a standard potassium feldspar (microcline)..............

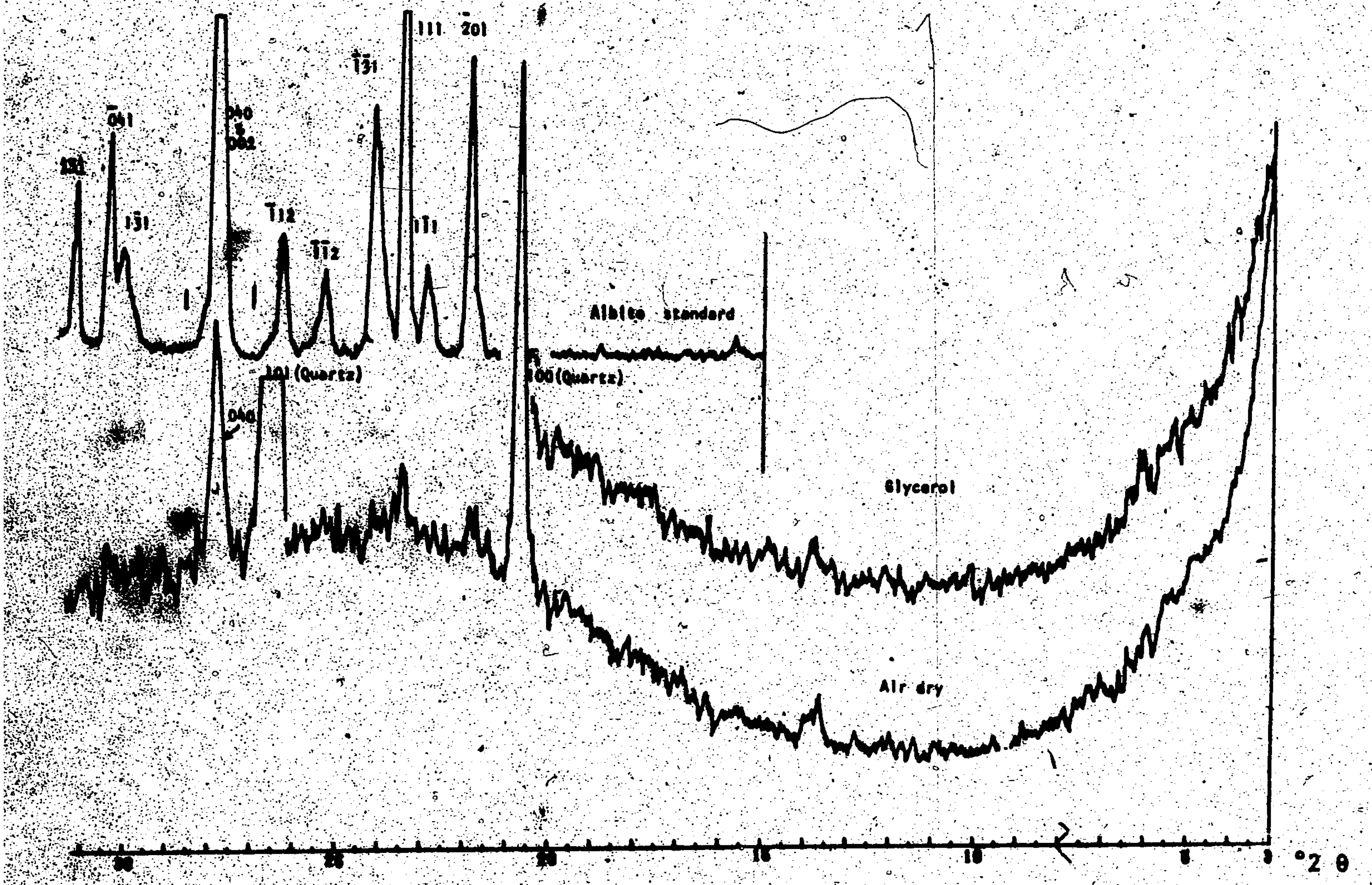


Figure 34. Typical X-ray diffractograms for the 2.59 to 2.72 g/cc separate from the coarse and medium silt fraction. The upper diffractogram represents a standard albite....

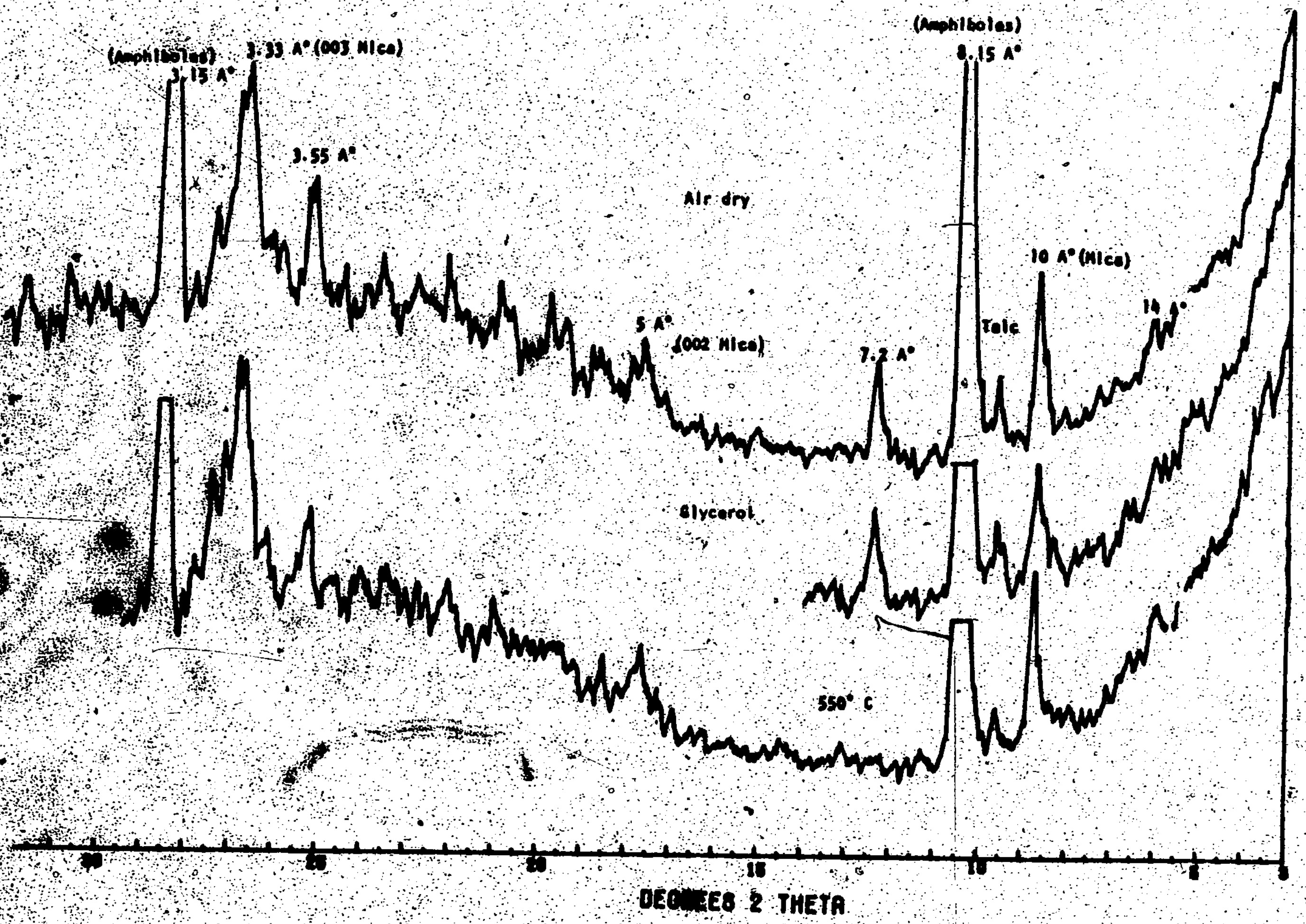


Figure 35. Typical X-ray diffractograms for the >2.72 g/cc specific gravity separate from the coarse and medium silt fraction

of Bayliss et al., 1970) are 50 to 70% amphiboles, 20 to 30% mica and 5 to 10% chlorite (with the possible presence of vermiculite). Other minerals are likely to be present, however, their quantities were too low to be detected by X-ray diffraction.

Bulk chemical analysis of the 2.53 to 2.59 g/cc and the 2.59 to 2.72 g/cc (Figures 36 to 38) substantiate the above general conclusion obtained from X-ray data which indicates potassim feldspars dominate in the first and plagioclases (after quartz) in the latter.

X-ray diffractograms of the 2.53 to 2.59 g/cc separates (Figure 39 and Figures 11 to 15, Appendix A) indicate the presence of both microcline and orthoclase polymorphs. The K-feldspar is likely perthitic in nature (as observed for the sand fraction) as suggested by the occurrence of the -201 peak at about 21.8° 2θ. The angular separation between the -201 (K-feldspar phase) and -201 (albite phase) is similar to those observed in X-ray diffratograms of the sand fraction and indicates a high degree of exsolution. The one shale sample generally shows the same peaks as those of soil horizon samples (Figure 39), which suggests that K-feldspar polymorphs in the soils are similar to those in this particular shale sample. Silt-sized feldspars in soil parent materials (of the study area) are mainly grains of shale origin admixed with feldspar grains from the shield which have been ground by glacial action to the silt size range.

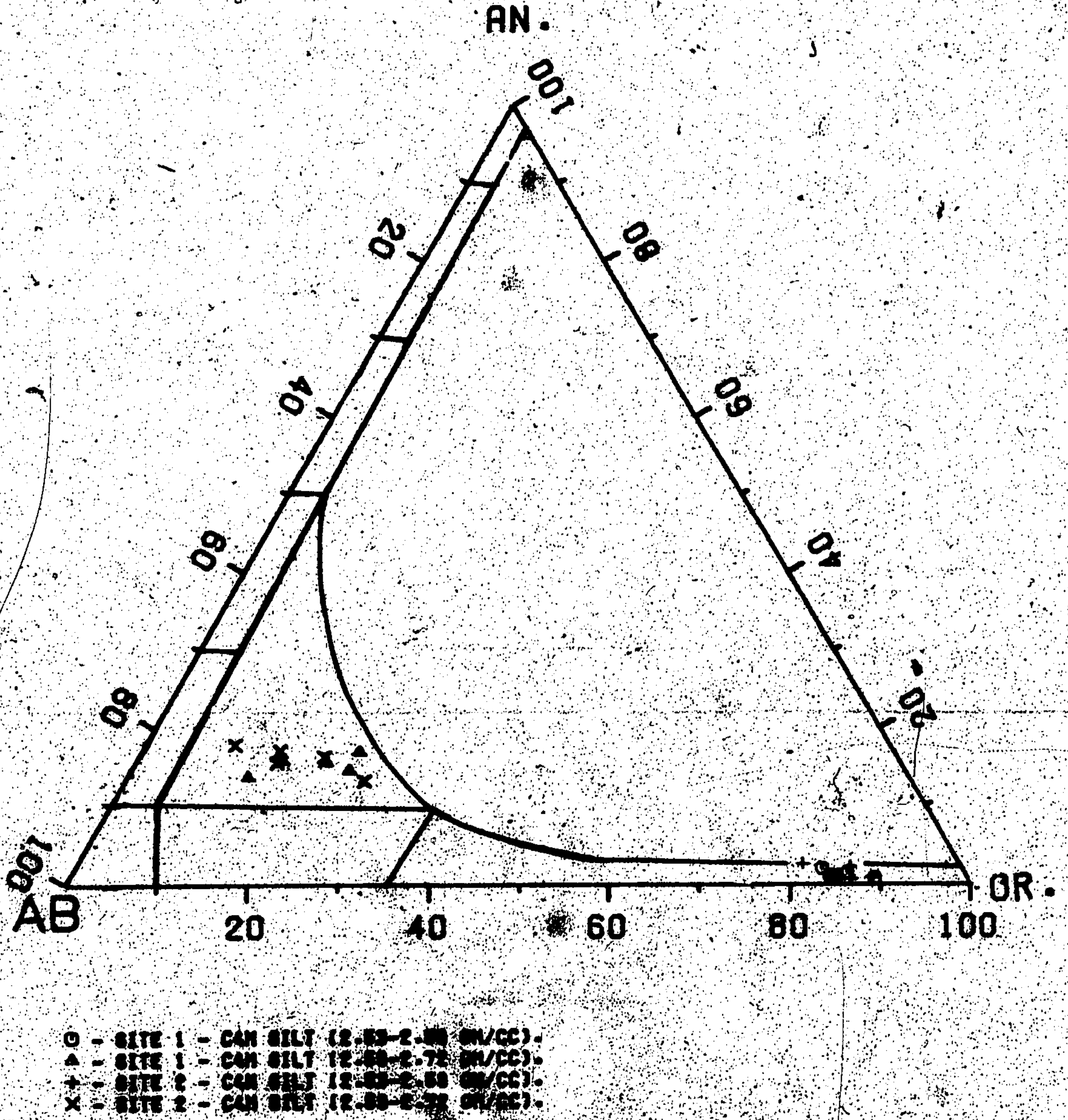


Figure 36. Composition of feldspars separated from all horizons from soils at sites 1 and 2.

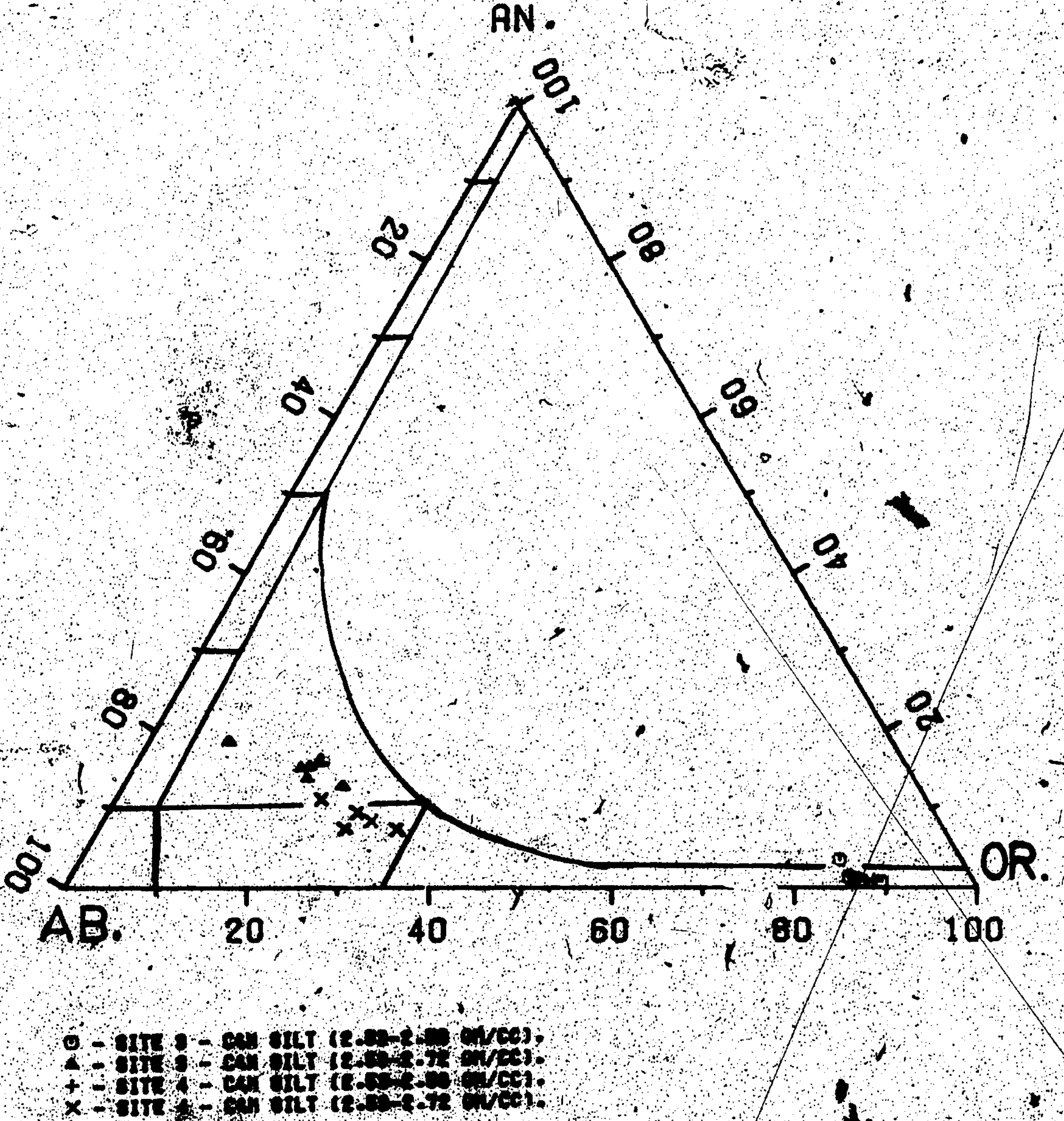


Figure 37. Composition of feldspars separated from silt horizons from soils at Sites 3 and 4.

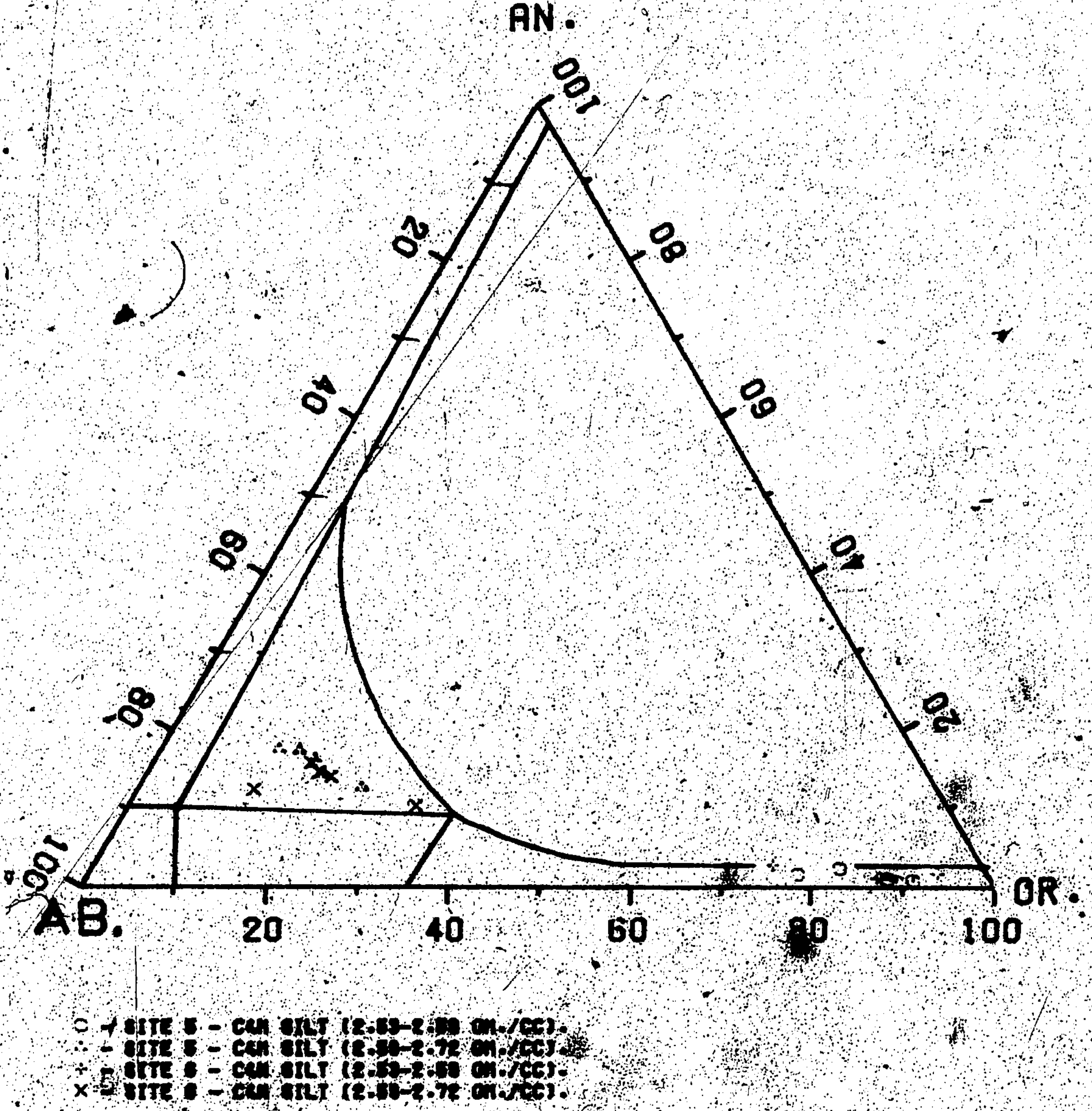


Figure 28. Composition of feldspars separated from all horizons from soils of [illegible] [illegible]

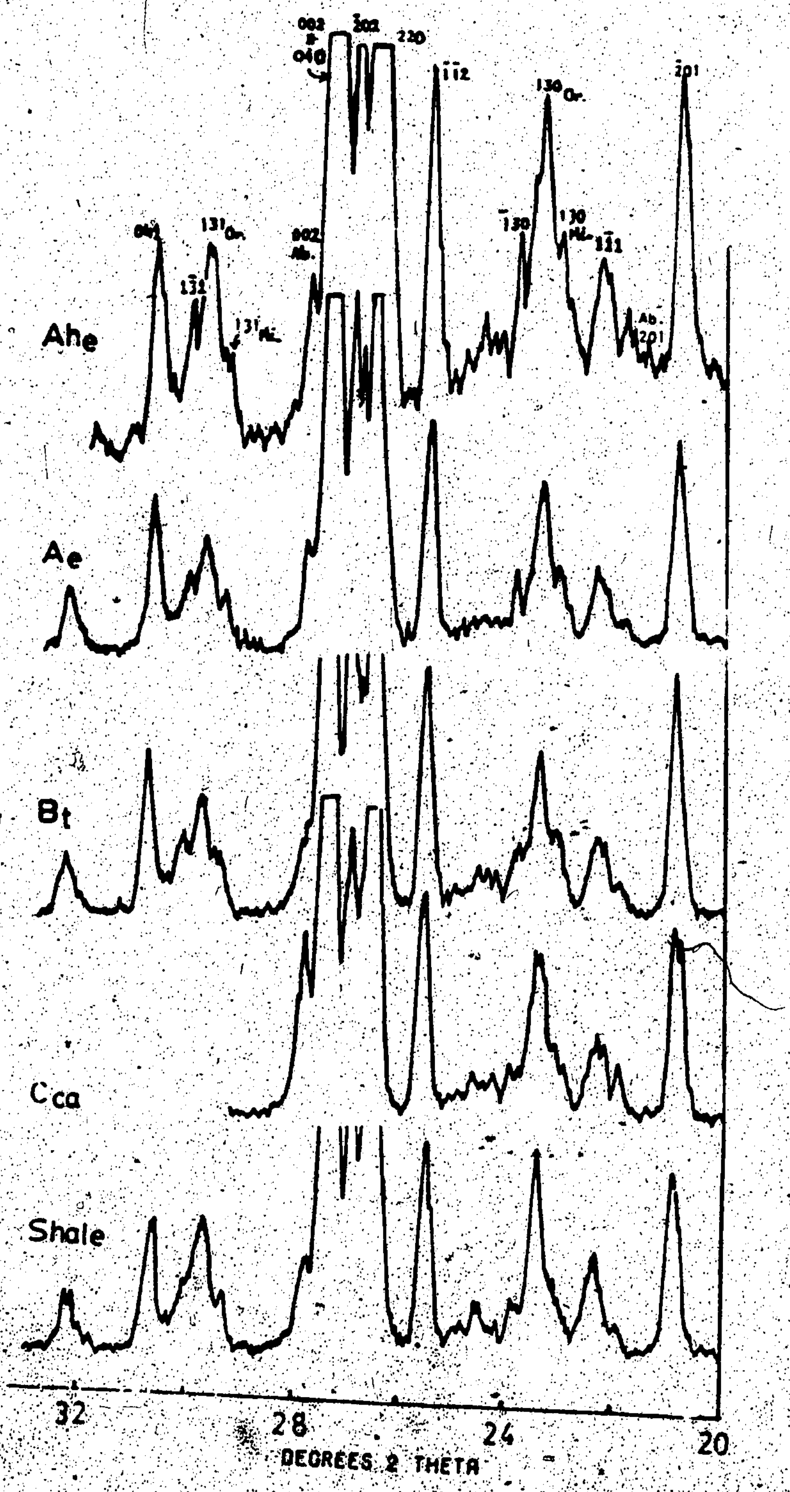


Figure 39. X-ray diffractograms for the 2.53 to 2.59 g/cc specific gravity separates from the coarse and medium silt fraction from the major genetic horizons of a Dark Gray Luvisol (site 2) and one shale sample from an outcrop in the sampling area (all indices except as noted are for K-feldspar).

Thus, the above results tend to suggest that silt-sized K-feldspar derived from shale bedrock is similar to those derived from granitic rocks. However, the previous statement is only a speculation because of the fact that only one shale sample was analyzed. More detailed feldspar mi[illegible]alogy of both shales and other possible source rocks such as granitic rocks, sandstones and siltstones should be studied before any reliable conclusions are drawn.

The average relative abundance of orthoclase in the K-feldspar fraction is 45 ± 10% using the data from all soil samples (Figure 40 and Table 12). The only apparent trend indicated by the relative amounts of the two potassium feldspar polymorphs within profiles is that surface horizons contain more microcline than do respective C horizons. The range of obliquity values for silt-sized microcline is 0.78 ± 0.07 to 0.91 ± 0.06 which is comparable to those observed in the fine sand fraction (average of 0.86 ± 0.06). Apparently, the nature of microcline in sedimentry rocks is similar to that in shield rocks because most of the feldspars in the sand are expected to be derived from shield materials while that of the silt from admixture of both shield and sedimentry rock materials. The magnitude of the obliquity values indicate the presence of highly ordered triclinic microcline in the K-feldspar separates.

Scanning electron micrographs of the potassium feldspar separates display similar features to those observed in the

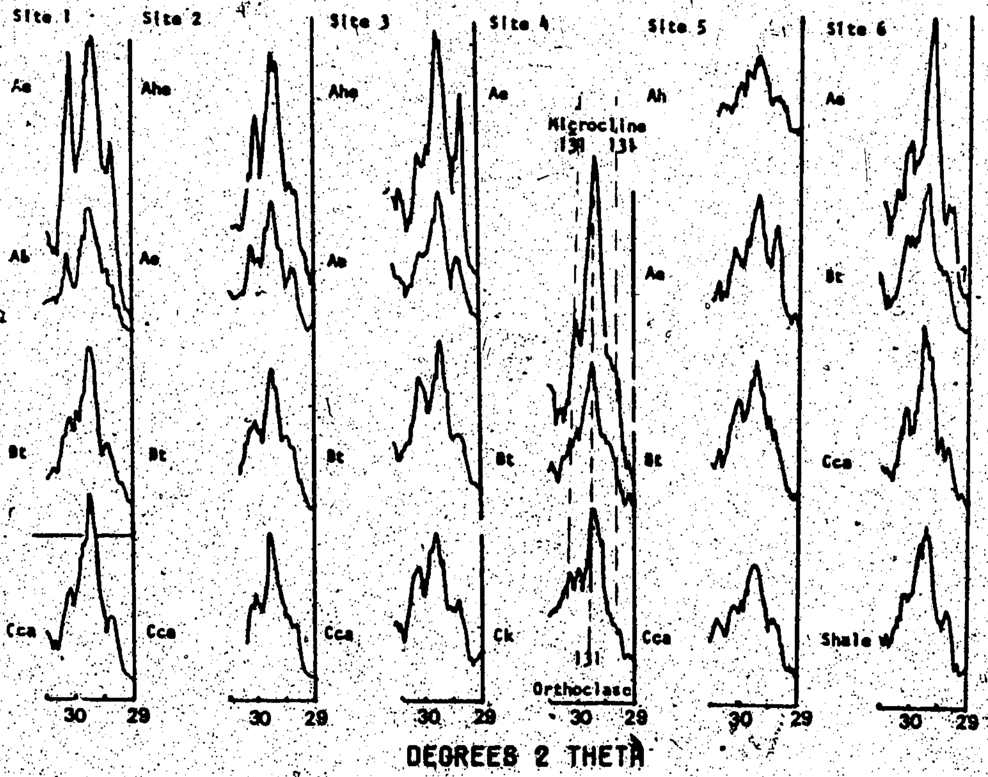


Figure 40. X-ray diffractograms showing variations in 131 and $\bar{1}31$ of microcline and 131 of orthoclase in the coarse and medium silt fractions.

Table 12. Relative proportions of orthoclase and microcline in K-feldspars and the obliquity value of microcline in the coarse and medium silt fraction (5-50 µm).

Soil	Horizon	Orthoclase%	Microcline%	Obliquity-Å
Orthic	Ae	34	66	0.91
Gray	AB	38	62	0.75
Luvisol	Bt	50	50	0.86
(site 1)	Cca	62	48	0.80
Dark	Ahe	49	51	0.86
Gray	Ae	40	60	0.86
Luvisol	Bt	41	59	0.74
(site 2)	Cca	52	48	0.98
Dark	Ahe	28	72	0.92
Gray	Ae	43	57	0.86
Luvisol	Bt	52	48	0.98
(site 3)	Cca	39	61	0.86
Orthic	Ae	56	44	0.83
Gray	Bt	37	63	0.81
Luvisol	Ck	60	40	0.70
(site 4)				
Eluviated	Ah	34	66	0.86
Black	Ae	26	74	0.88
Chernozemic	Bt	34	66	0.75
(site 5)	Cca	41	59	0.86
Orthic	Ae	53	47	0.93
Gray	Bt	50	50	0.80
Luvisol	Cca	56	44	0.92
(site 6)				
Shale		50	50	0.92
Grand mean		45	55	0.86

sand fraction and include solution pitting and secondary mineral formation (Plate 10). Despite the relatively large specific surface of silt particles as compared to sand particles some grains still display a fresh surface appearance. The presence of both weathered and relatively unaltered feldspar grains in the soil solum as well as in C horizons suggests the weathered grains are largely an inherent feature of the till and are not solely the result of pedogenic weathering.

X-ray diffractograms of the 2.59 to 2.72 g/cc separate (Figure 41) and selected representative diffractograms (Figures 16 to 18, Appendix A) indicate quartz and plagioclase dominate this fraction. The peak intensities of plagioclase minerals are appreciably higher and more defined than those in the fine sand fraction, a result of the higher percentage of plagioclases in the silt fraction relative to those of the sand fraction. The 130 reflection of K-feldspar is present in all samples with an exceptionally high relative intensity in some samples (Figure 41 and Figures 16 to 18, Appendix A). The presence of variable amounts of K-feldspar in plagioclase separates necessitated recalculating the chemical data so that the 2.59 to 2.72 g/cc separate could be expressed solely in terms of Na-Ca feldspar composition (Table 14, Appendix A).

Quantitative mineralogical data for the coarse and medium silt fraction was obtained in the same manner as for

Description of Plate 10

A to C: Micrographs showing K-feldspar grains displaying different degrees of alteration. Micrograph A illustrates the appearance of a relatively unaltered feldspar grain. Although some breakage of the grains has occurred, preservation of sharp edges indicates minimal physical abrasion during transport. Solution pitting and the rounding of edges of some feldspar grains (as in micrograph B) indicates both chemical and physical weathering has occurred. In micrograph C, stacks of plate-like structures on the surface of the feldspar grain suggests the presence of phyllosilicate material.

D to E: Plagioclase feldspar grains showing pronounced physical and chemical alteration typical of this feldspar fraction. Plagioclases always appeared more weathered than K-feldspars. The spongy appearance of plagioclase feldspar (e.g. micrograph E) indicates extensive dissolution through chemical action. The presence of secondary minerals on the surface of the feldspar grain (micrograph F) is evident from numerous thin curled platy structures.

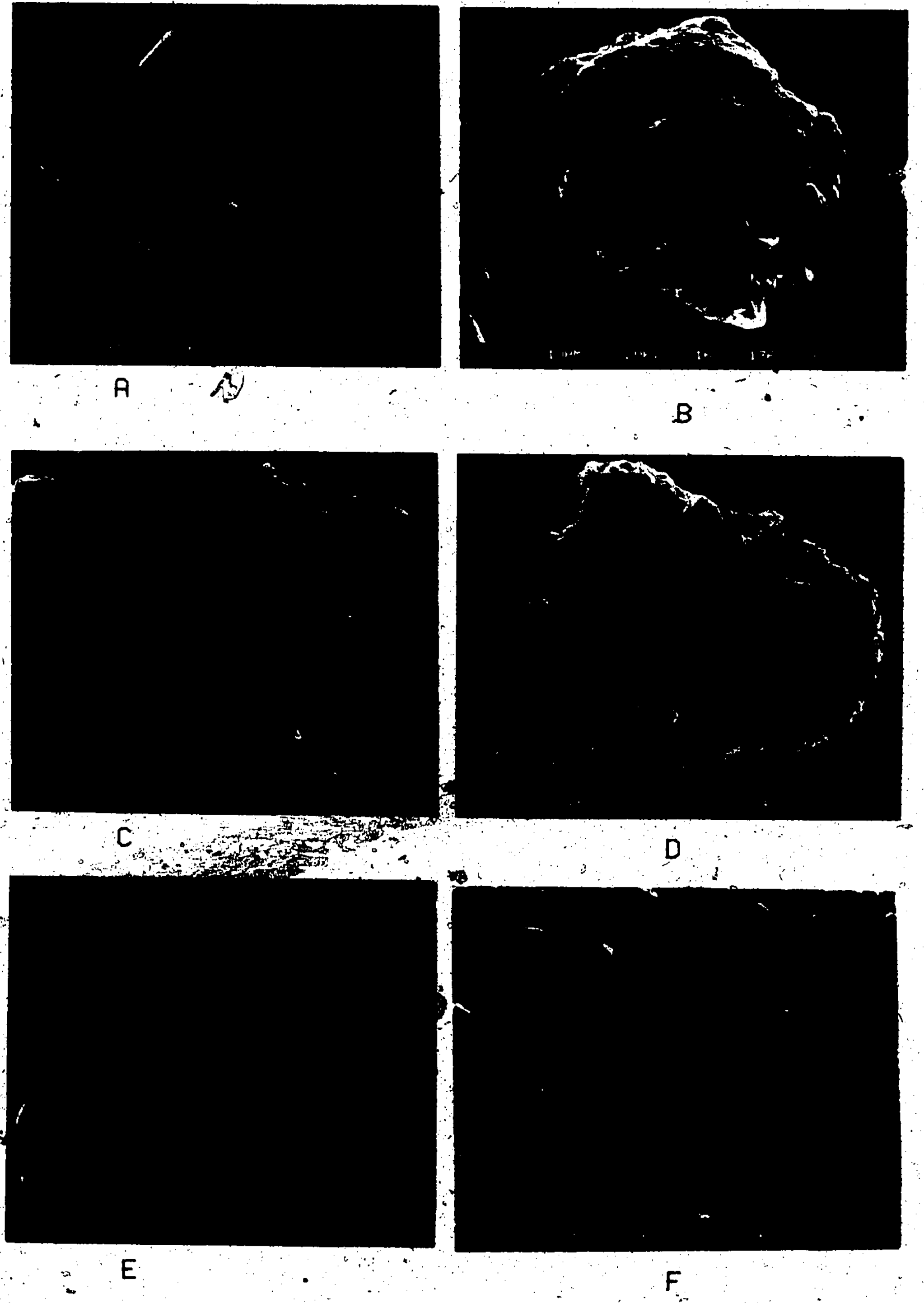


Plate 10. Scanning electron micrographs of several typical silt-sized feldspar grains. Micrographs A to C are for alkifeldspars (2.53 to 2.59 g/cc separate) while D to F are for plagioclases (2.59 to 2.72 g/cc separate).

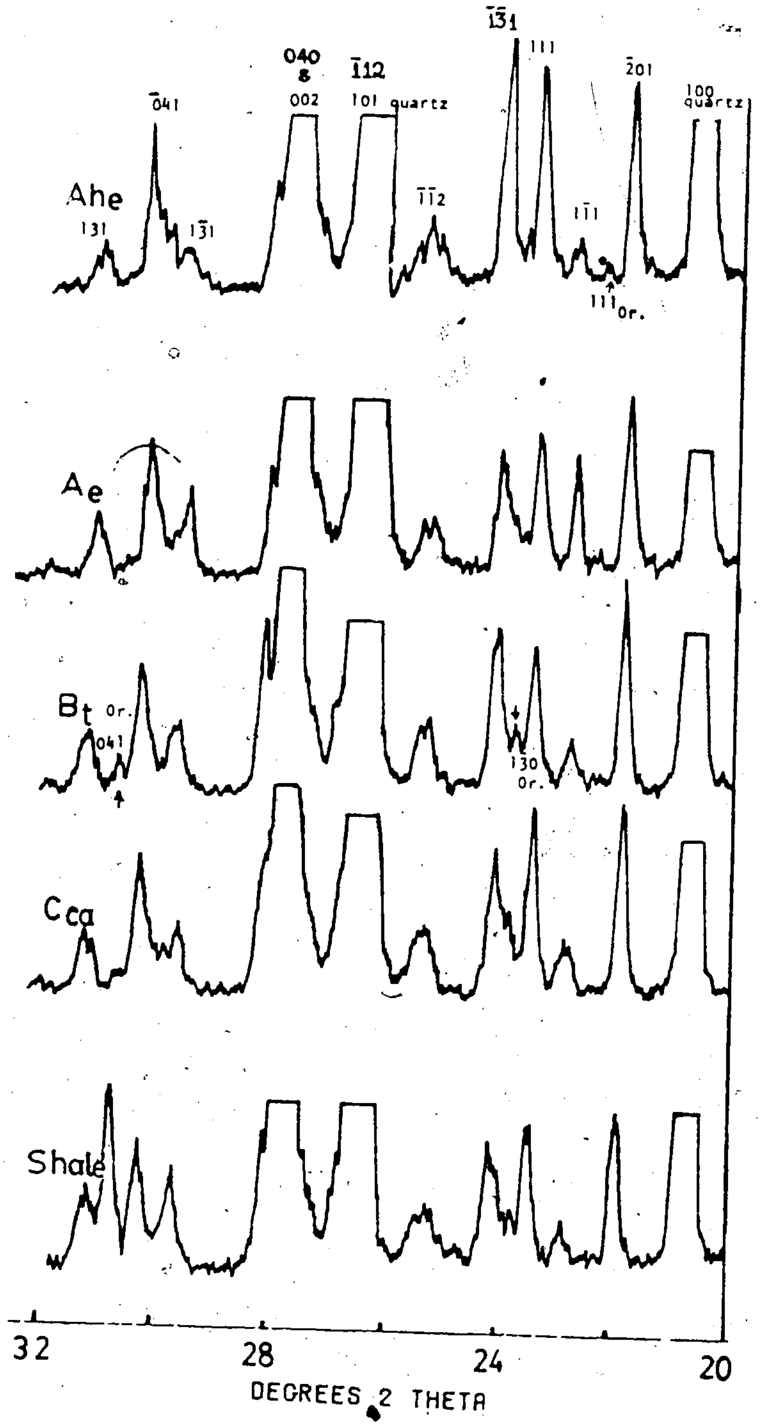


Figure 41. X-ray diffractograms for the 2.59 to 2.72 g/cc specific gravity separates from the coarse and medium silt fraction from the major genetic horizons of a Dark Gray Luvisol (site 3) and for one shale sample from an outcrop in the sampling area (all indices except as noted are for albite...............

the fine sand fraction. In a general sense, composition of silt fractions are similar for all soils in that quartz is always the dominant mineral constituent with plagioclases next in abundance followed by K-feldspars and heavy minerals (Figures 42 to 47). Mineral composition of the coarse and medium silt fractions from sites 1, 2, 3 and 5 appear quite similar to each other when comparing respective horizons while all samples from sites 4 and 6 contain higher amounts of quartz than is present in samples from the other four sites. There is a concomitantly lower content of feldspars in silt from sites 4 and 6 as compared to amounts of feldspars in samples from sites 1, 2, 3 and 5. Parent materials of profiles 1, 2, 3 and 5 contain an average of 11.7% K-feldspar and 18.6% Na-Ca feldspars versus 9.1% K-feldspar and 12.9% Na-Ca feldspars in parent materials at sites 4 and 6. The variations in amounts of mineral constituents among the various sites could be due to mineralogical differences of source rock material (especially underlying shales), nearness to Shield rock and differences in the amount of Shield rock incorporated into the till deposits.

d. Fine silt fraction

Fine silt constitutes 1.5 to 7.1% of the total soil by weight. This fraction was characterized by X-ray diffraction only (Figures 48 to 50) and thus estimates of relative mineral abundances are semiquantitative (Tables 13 and 14). These estimates were determined (except for vermiculite)

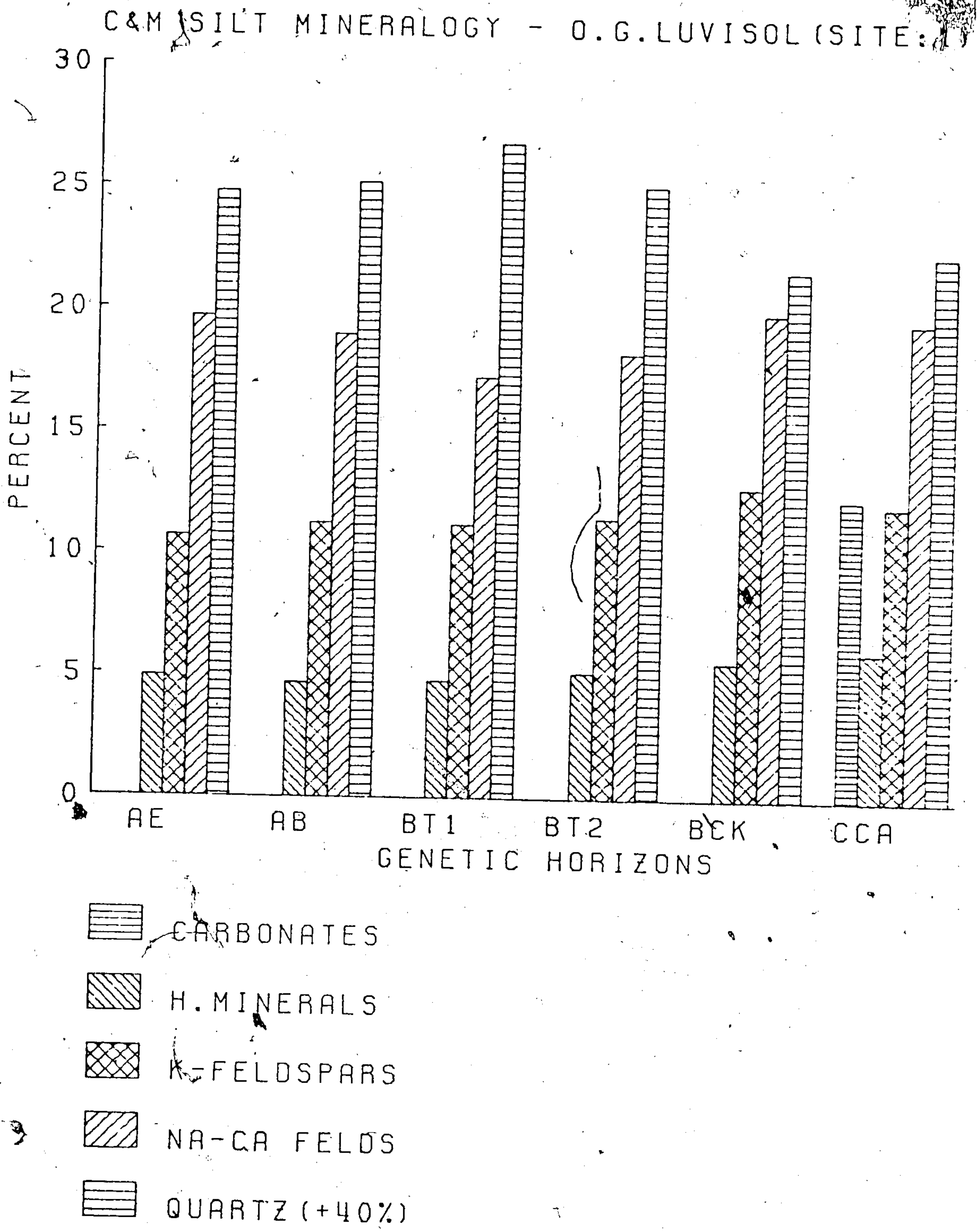


Figure 42. Histograms showing the mineralogical composition of the coarse and medium silt fractions (site 1)...........................

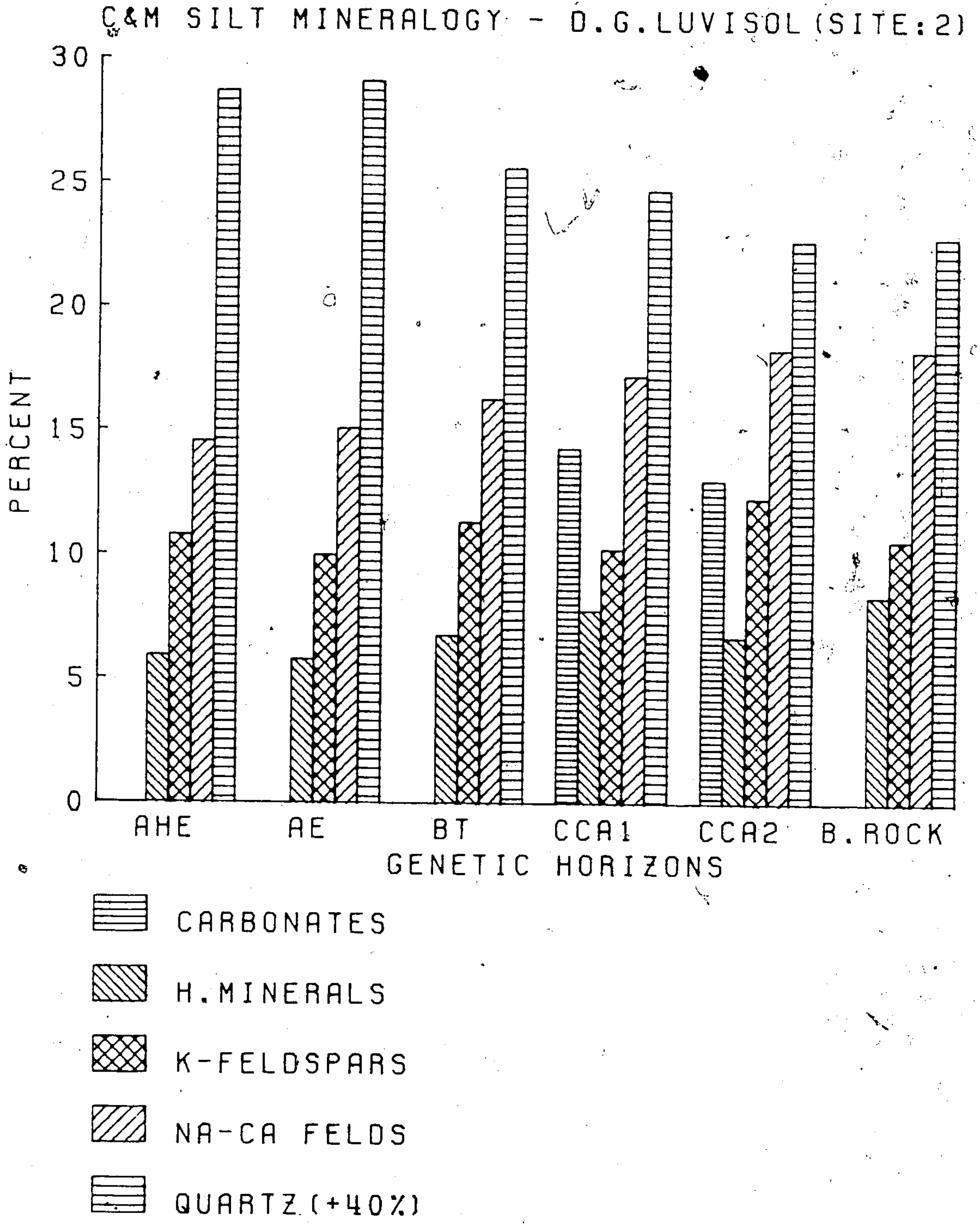


Figure 43. Histograms showing the mineralogical composition of the coarse and medium silt fractions (site 2)............................

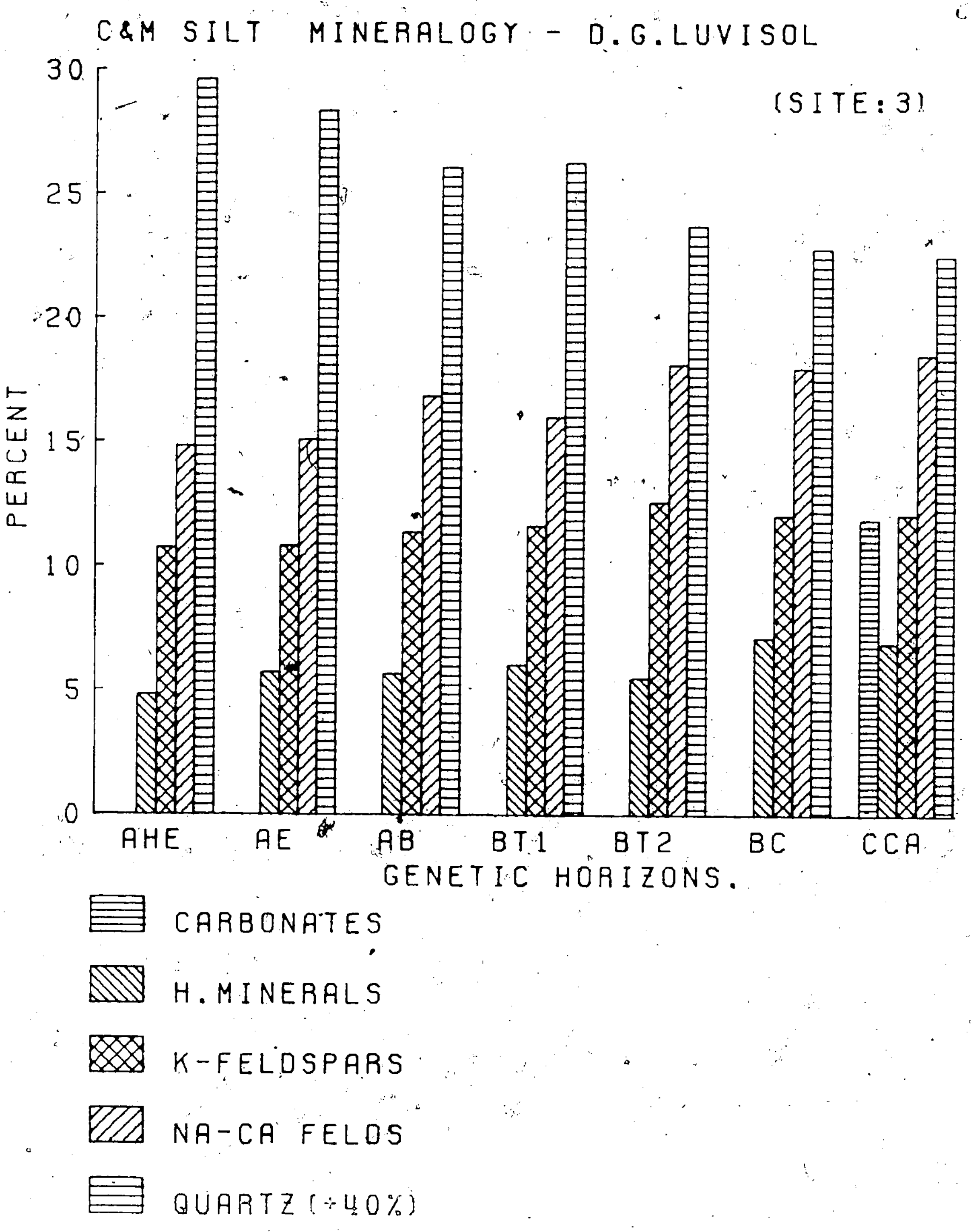


Figure 44. Histograms showing the mineralogical composition of the coarse and medium silt fractions (site 3)...........................

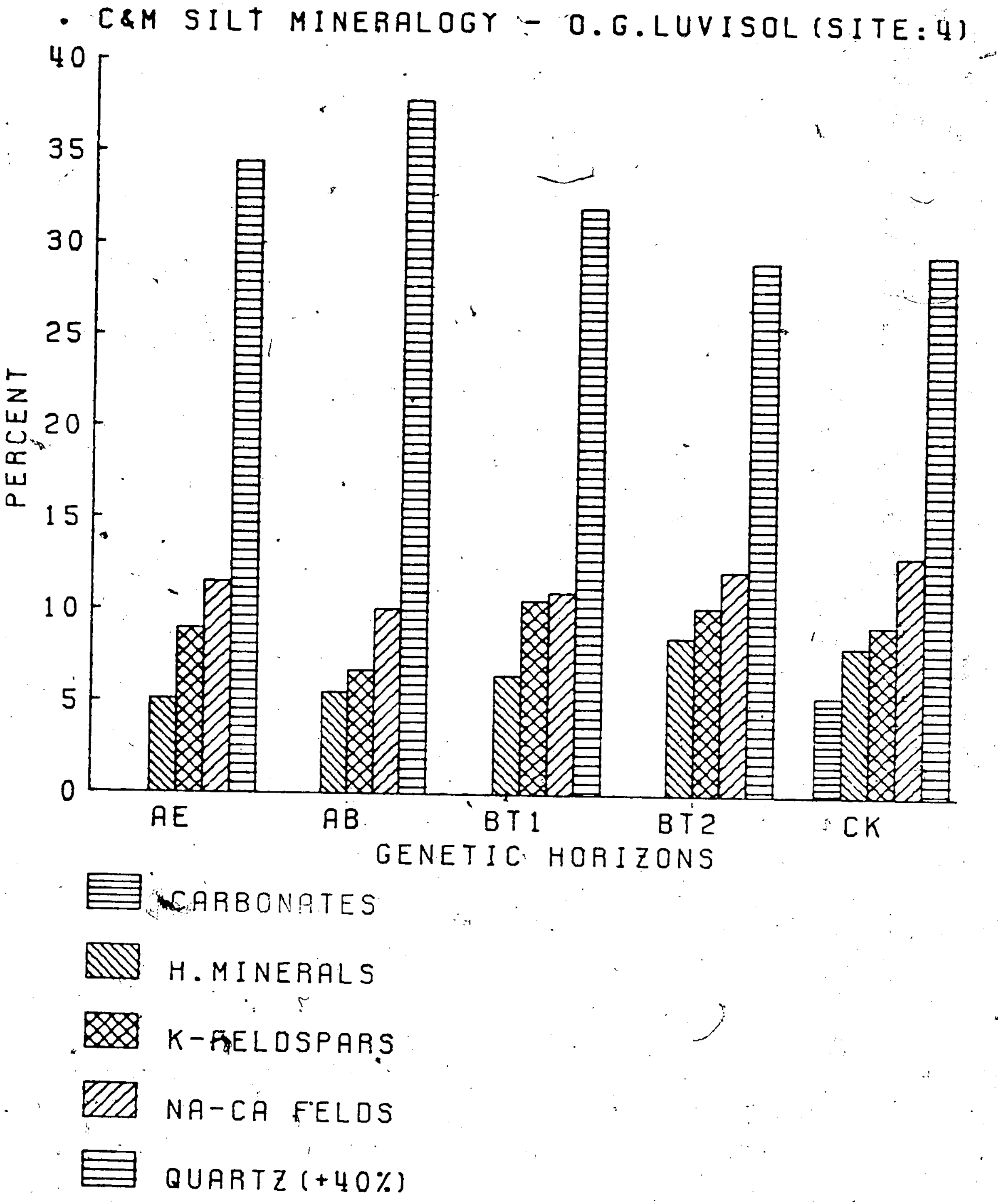


Figure 45. Histograms showing the mineralogical composition of the coarse and medium silt fractions (site 4)

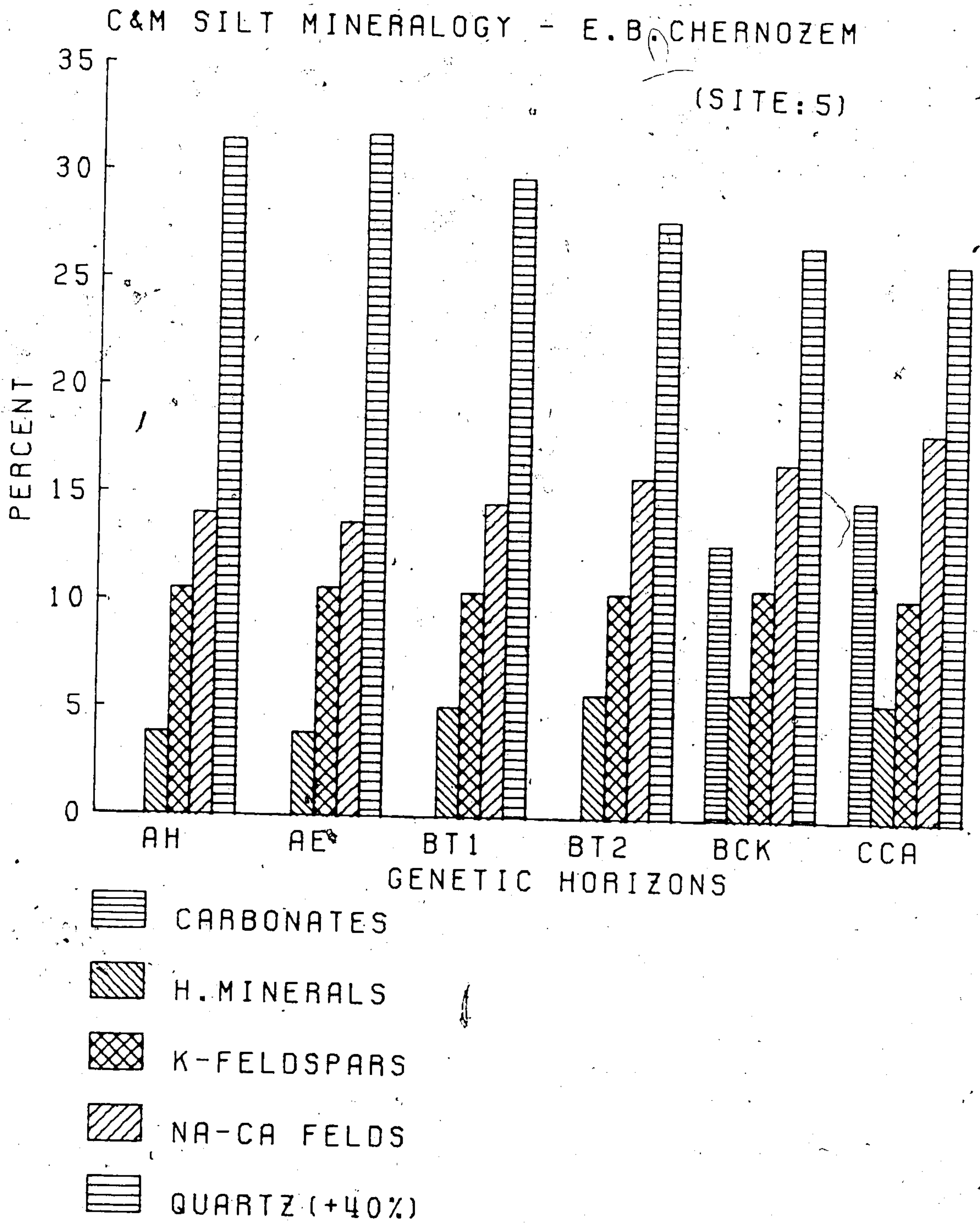


Figure 46. Histograms showing the mineralogical composition of the coarse and medium silt fractions (site 5)..............................

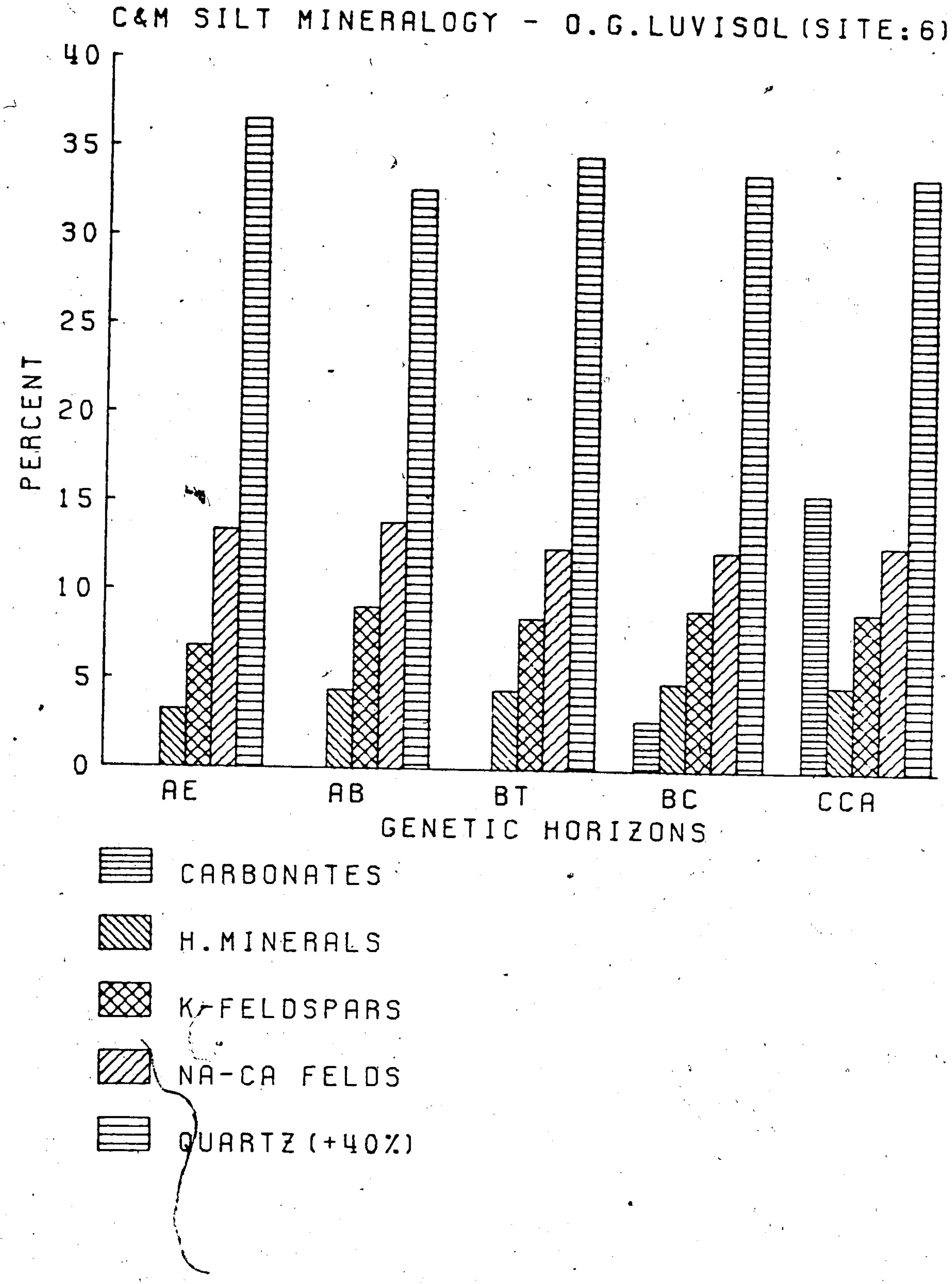


Figure 47. Histograms showing the mineralogical composition of the coarse and medium silt fractions (site 6)..........................

using corrected and normalized intensity ratios as described by Bayliss et al. (1970). The reflections and factors used to multiply their intensities are 14 Å for chlorite by 17, 10 Å for mica by 70, 8.5 Å for amphiboles by 84, 4.26 Å for quartz by 84, 3.24 Å for potassium feldspar by 45 and 3.19 Å for plagioclase by 30 after which the sum of all factors multiplied by intensities is divided into each factor multiplied by its intensity to arrive at the relative quantities. The following mineral abundances were determined: quartz (28 to 51%), mica (12 to 27%), kaolinite (9 to 21%), Na-Ca feldspar (4 to 11%), potassium feldspar (3 to 8%), amphiboles (1 to 18%), vermiculite (1 to 20%) and chlorite (<1 to 2%).

Variation in mineral distribution within profiles appears small possibly due to the semiquantitative nature of the data. Variations between profiles appear to be present in content of amphiboles, K-feldspars and Na-Ca feldspars. Using this semiquantitative data it seems that content of amphiboles at site 4 is the lowest when compared to all other profiles (1 to 3% at site 4 compared to 5 to 19% at sites 1, 2, 3 and 5). Site 4 and 6 also have the lowest content of K-feldspar and Na-Ca feldspar. These results agree with data obtained for the coarse and medium silt where the lowest content of feldspar occurred at site 4 and 6 (the two sites furthest away from the remaining four sites and the Shield).

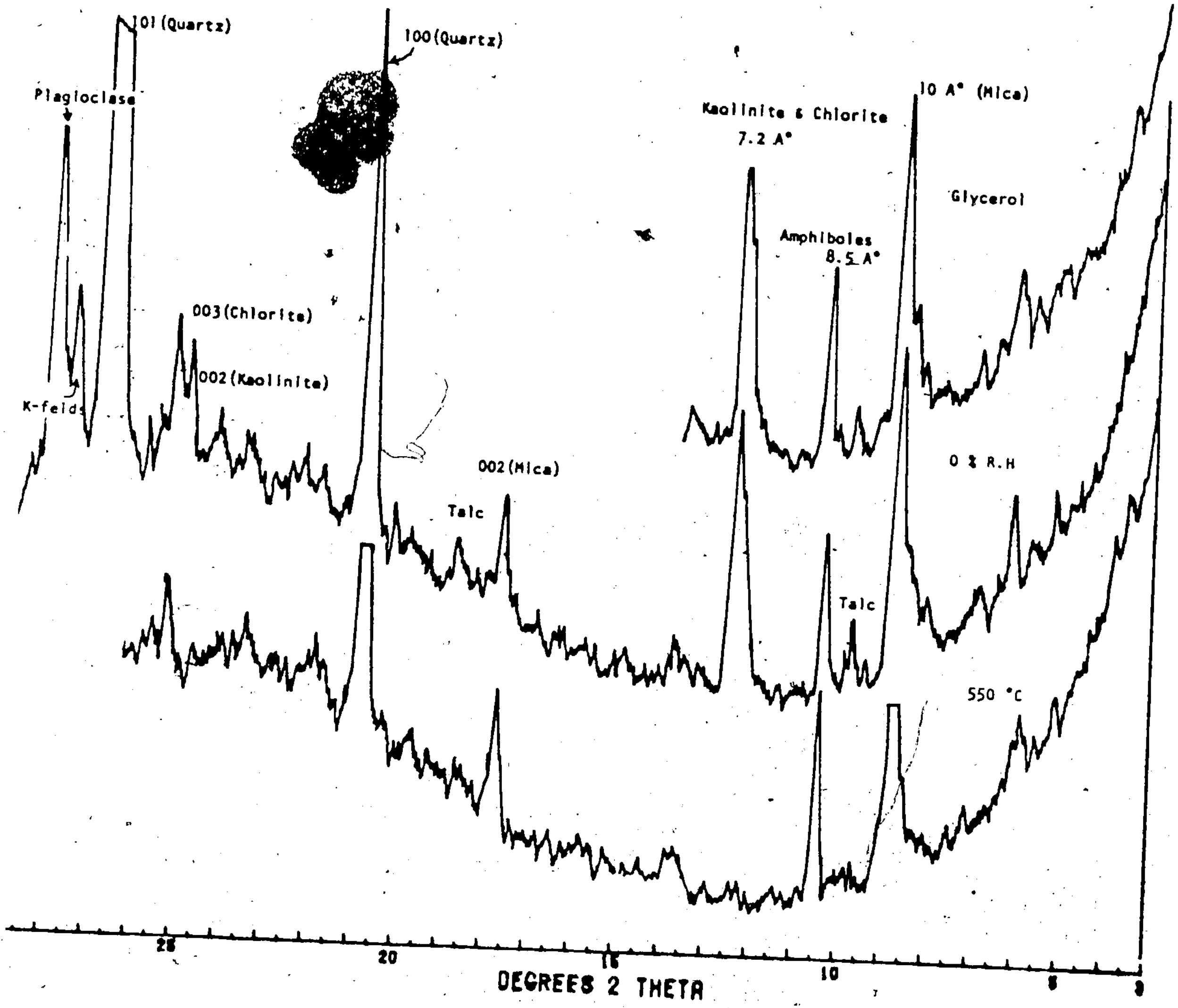


Figure 48. X-ray diffractograms for the fine silt fraction separated from the Ae horizon from an Orthic Gray Luvisol (site 1)..............

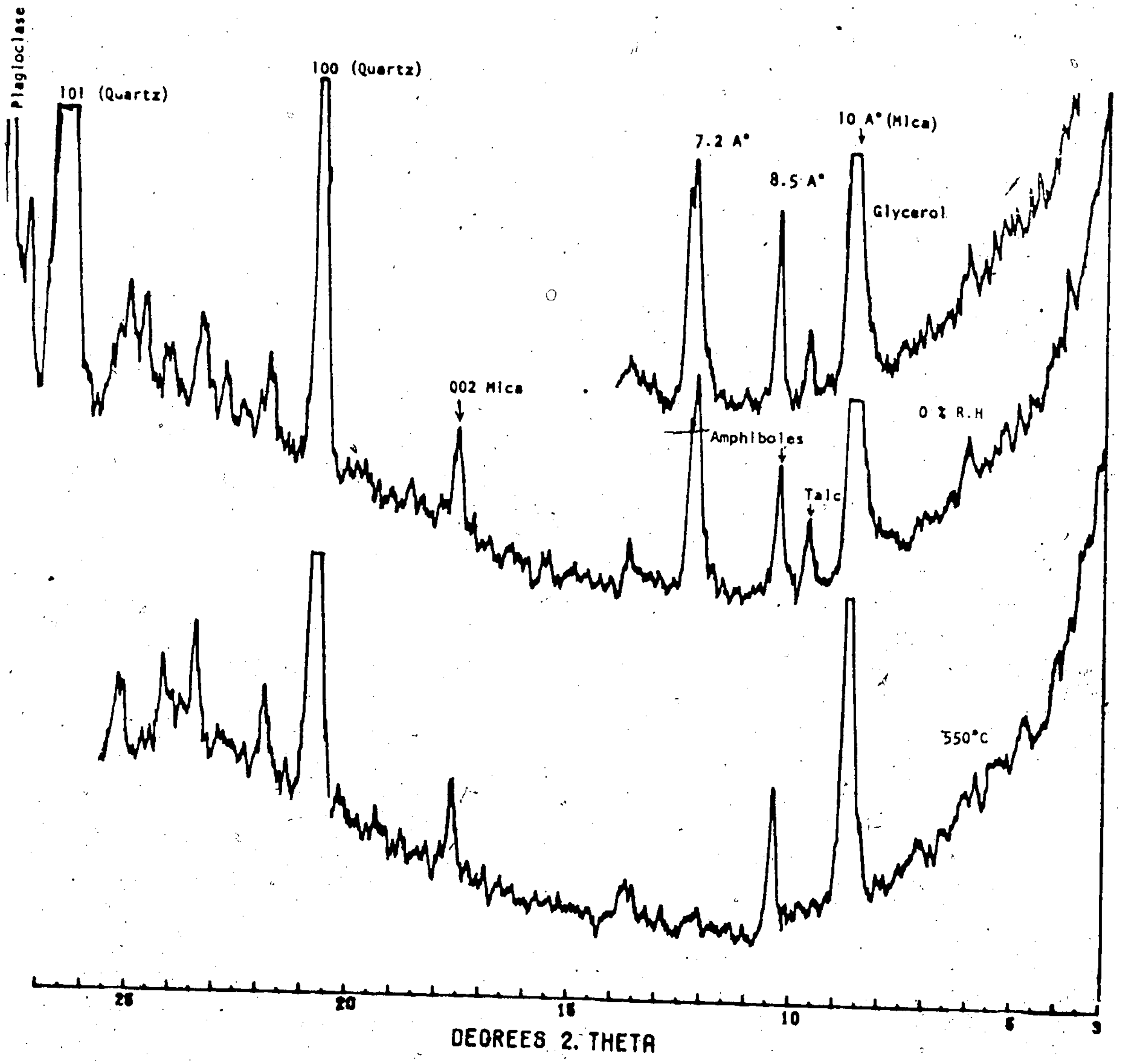


Figure 49. X-ray diffractograms for the fine silt fraction separated from the Bt horizon from an Orthic Gray Luvisol (site 1)..............

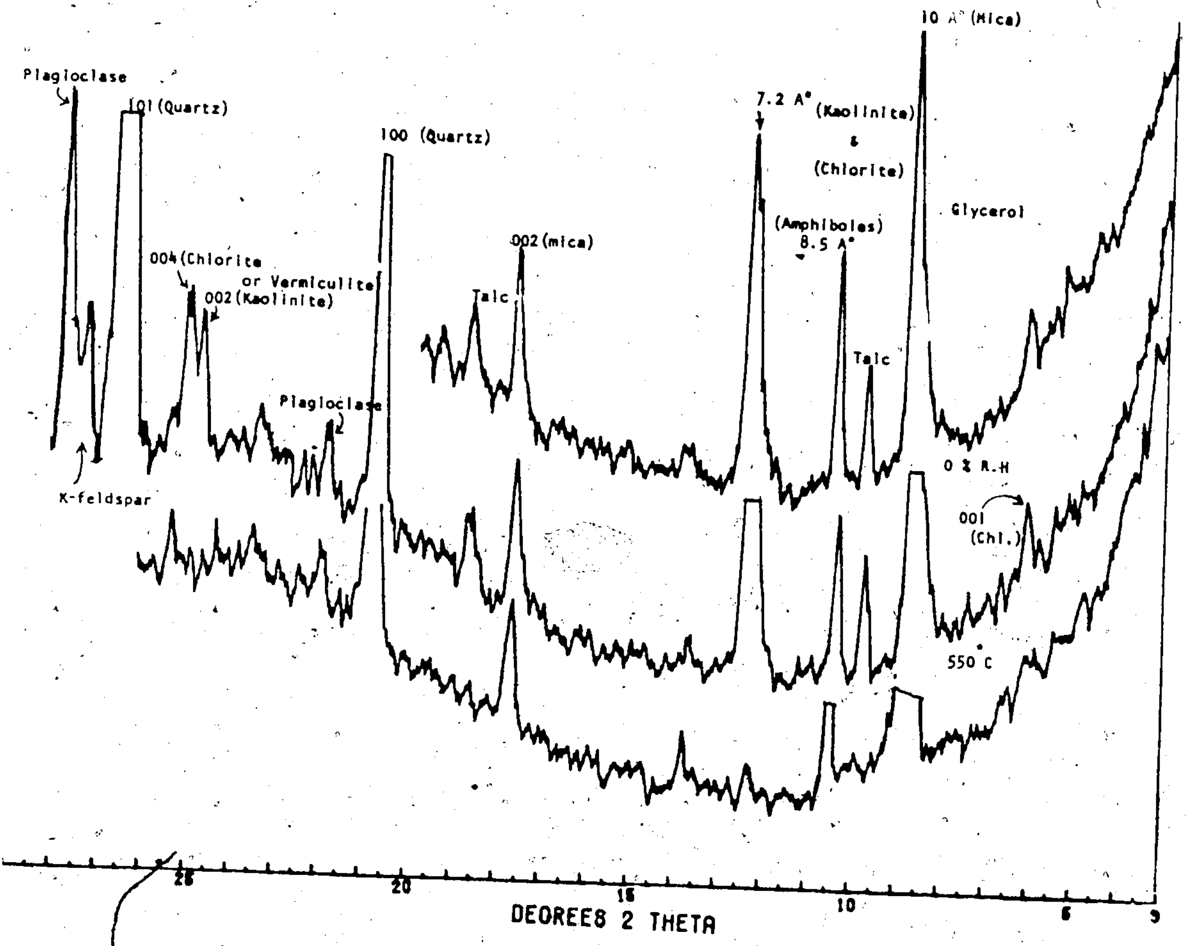


Figure 50. X-ray diffractograms for the fine silt fraction separated from the Cca horizon from an Orthic Gray Luvisol (site 1)..............

Table 13. Relative proportions of various minerals in fine silt fractions (sites 1 to 3).

Horizon	Qtz	Na-CaF	Mi.	KF	Kaol.	Amph.	Verm.	Chl.
				Relative content %				
			Orthic Gray Luvisol (site 1)					
Ae	34	9	18	8	16	13	1*	1
AB	32	6	27	7	15	9	3	1
Bt	42	9	20	6	10	10	2	1
Cca	35	9	16	6	14	14	3	1
			Dark Gray Luvisol (site 2)					
Ahe	38	7	16	6	13	5	13	2
Ae	41	7	17	6	16	9	3	1
Bt	44	9	14	6	13	8	5	1
Cca	37	10	12	6	14	18	3	1
			Dark Gray Luvisol (site 3)					
Ahe	51	7	18	4	9	5	6	<0.5
Ae	45	7	18	7	13	7	2	1
Bt	33	9	19	8	15	10	5	1
Cca	42	11	15	7	13	8	3	1

* Values recalculated from those determined by KEC and CaEC for the total silt fraction since most vermiculite occurs in the fine silt versus the coarse plus medium silt fraction as noted from XRD patterns.

Abbreviations: Qtz = quartz, Na-CaF = NA-Ca feldspars,
KF = K-feldspars, Kaol = kaolinite,
Amph. = amphiboles, Verm. = vermiculite,
Chl. = chlorite, Mi. = mica

Table 14. Relative proportions of various minerals in fine silt fractions (sites 4 to 6).

	Relative content %							
Horizon	Qtz	Na-CaF	Mi	KF	Kaol	Amph.	Verm.	Chl.
		Orthic Gray Luvisol (site 4)						
Ae	44	5	20	6	15	2	6*	2
AB	41	5	22	5	14	3	8	2
Bt	49	6	17	4	14	1	8	1
CK	45	5	21	3	16	3	6	1
		Eluviated Black Chernozemic (site 5)						
Ah	37	6	13	6	11	5	20	2
Ae	35	6	20	9	17	8	4	1
Bt	38	7	19	8	16	8	3	1
Cca	28	8	21	6	20	14	2	1
		Orthic Gray Luvisol (site 6)						
Ae	36	4	20	6	21	6	5	2
AB	37	5	20	6	19	6	6	1
Bt	39	5	19	5	20	5	6	1
Cca	40	6	21	6	20	7	1	1

* Values recalculated from those determined by KEC and CaEC for the total silt fraction since most vermiculite occurs in the fine silt versus the coarse plus medium silt fraction as noted from XRD patterns.

Abbreviations: Qtz = quartz , Na-CaF= Na-Ca feldspars,
KF = K-feldspars, Kaol = kaolinite,
Amph. = amphiboles , Verm. = vermiculite,
Chl. = chlorite , Mi. = mica,

e. Clay mineralogical characterization

The mineralogy of the clay fraction was studied in detail by X-ray diffraction, total dissolution, cation exchange capacity and surface area characteristics. These methods enable semi-quantitative and quantitative estimates to be made of the different clay mineral species which can not be done from X-ray diffraction data alone.

A minimum of six cation saturation/solvation/hydration/heat treatments are needed to ensure proper identification of clay minerals and clay mineral intergrades by X-ray diffraction techniques. Calcium saturation and solvation with glycerol for example, aids in identification of vermiculite (14.3 Å) and smectite (16.6 to 17 Å) (Harward and Brindly, 1964). Chlorite and chloritic intergrades are distinquished by the 300° and 500° C heat treatments of K-saturated specimens; chloritic intergrades give a "shoulder" or peak at 10 to 13 Å after the 300° C heat treatment. K-saturated clay samples analyzed at known relative humidities are necessary to verify the presence of vermiculite which does not rehydrate whereas smectite species do, producing a peak between 12-13 Å with concomitant decrease in the 10 Å peak (Sayegh et al., 1965; Harward et al., 1969).

The mica content in the clay fraction was calculated from the K20 values obtained by chemical analysis, assuming the micaceous minerals contained approximately 10% K20

(Alexiades and Jackson, 1966). Where X-ray diffraction patterns indicated the presence of feldspars, the amounts of K-feldspars were estimated from the intensity of it's -201 peak at about 3.24 Å using the method of Bayliss et al. (1970). Determination of the amounts of potassium feldspars in the clay fraction makes it possible to assign the remaining amount of K20 to mica.

Smectite was determined from cation exchange capacity (Ca^{++}EC) after deducting 159 me/100 g for the vermiculite present (Alexiades and Jackson, 1965) and assuming an average cation exchange capacity of 105 me/100 g for smectite (Alexiades and Jackson, 1966). In the calculation of smectite, no direct correction was made for depotassified mica because it is included in the value reported as "vermiculite". Surface area measurements (after deducting 810 m2/g for vermiculite and 20 m2/g for mica) was used to estimate smectite assuming a surface area of 810 m2/g for smectite (Carter et al., 1965). Surface area of vermiculite was considered to be 810 m2/g rather than the 350 m2/g of Carter et al. (1965) because vermiculite present in the soil, unlike geological specimens of vermiculite, usually expands to 17 Å when equilibrated with ethylene glycol. Surface area of expanding 2:1 layer silicates including montmorillonite and vermiculite should be 808 m2/g for particles in the range of 2 to 0.2 micrometer and 813 m2/g for particles in the range of 0.2 to 0.08 micrometer (Jackson, 1975) rather than the higher values sometimes reported in the literature. In most instances, content of smectite based on the cation exchange capacity method is larger than contents based on surface area measurements. The greatest descrepancy between the two methods of determining smectite occurs for clays from surface A horizons (particularly for Ah and Ahe samples). For clay separates from B and C horizons, the content of smectite based on surface area method tends to be the same or slightly higher than the content arrived at by cation exchange capacity measurements. Clay fractions from surface A horizons likely contain substantial amounts of organic substances even after peroxide treatment as has been demonstrated by Dudas and Pawluk (1970). These unoxidized organic materials likely contribute to the exchange capacity and result in an overestimation of smectite. As a consequence, smectite values based on the surface area technique for clay separates from Ah and Ahe horizons are considered to represent contents of smectite in these horizons. For other soil horizons, smectite contents were based on the cation exchange capacity method since it is generally considered that the cation exchange capacity method is superior to the surface area method in measuring the amount of smectite in organic matter free clay samples (Borchardt, 1977).

Vermiculite was determined from the difference between calcium and potassium exchange capacity and an assumed vermiculite interlayer exchange capacity of 154 me/100 g (Coffman and Fanning, 1974).

Relative amounts of kaolinite, chlorite, quartz, potassium feldspars and plagioclase feldspars were semi-quantitatively estimated in the remainder of the clay fraction [100% - (% mica + % vermiculite + % smectite)] using corrected and normalized intensity ratios of characteristic X-ray diffraction peaks. The correction factors used to multiply the intensities of X-ray diffraction reflections were 17 for chlorite, 70 for kaolinite, 84 for quartz, 45 for potash feldspar and 30 for plagioclase according to procedures outlined by Bayliss et al. (1970).

X-ray diffraction data of random powder clay samples was used to determine whether the clay minerals were dioctahedral or trioctahedral. The location of the 060 peaks of phyllosilicate minerals depends on the amount of octahedral substitution present. Diffraction peaks near 61.85 °2θ (1.500 Å) are characteristic of dioctahedral structures while those near 60.2 to 61.0 °2θ (1.537 to 1.519 Å) are characteristic of trioctahedral structures (Douglas, 1977). Diffraction peaks from quartz were found not to interfere with the identification of dioctahedral and trioctahedral phyllosilicates when specimens were heated to 550°C (Figure 51). Also, heating specimens to 550°C eliminated the possible interference from kaolinite (Figure 51).

X-ray diffraction patterns of clay ($\leq$2 micrometer) from

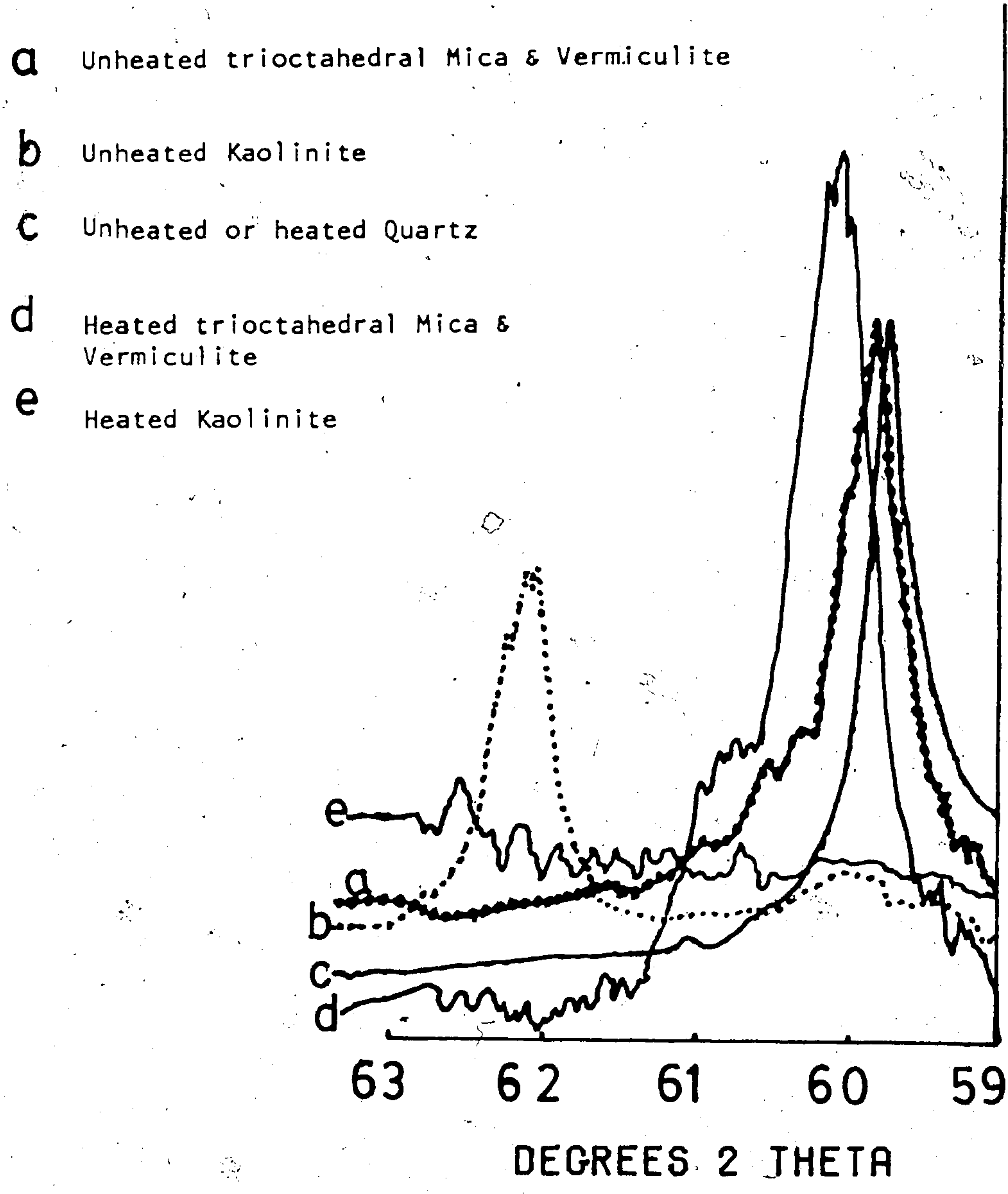


Figure 51. X-ray diffractograms showing peak positions for trioctahedral minerals, kaolinite and quartz before and after heating................

the main genetic horizons of two profiles (site 1 and 2) are included in Figures 52 to 58 to illustrate diffraction behavior and method of interpretation. X-ray diffractograms for the major horizons of the remaining profiles and one shale sample from the Sand River area can be found in the Appendix (Figures 19 to 33). Physical and chemical analysis of these clays together with their estimated or determined amounts are presented in Tables 15 to 20.

On X-ray diffractograms, the presence of smectites and/or soil vermiculite is indicated by the occurrence of peaks near 15 Å for the Ca-saturated specimen at 54% relative humidity. A peak near 17 Å for the ethylene glycol solvated sample is indicative of smectites, however, some soil vermiculites may also expand with ethylene glycol. The occurrence of a well defined 14 Å peak along with a 17 Å peak for glycerol-treated samples indicates the presence of both vermiculite (and/or chlorite) and montmorillonite. The glycerol molecule being less polar than the ethylene glycol molecule does not expand vermiculite. The presence of vermiculite is only confirmed by assessing the degree of rehydration by examining the 10 Å and 14 Å peak intensities of K-saturated specimens when equilibrated at 0% and at 54% relative humidity. Dehydrated K-saturated vermiculites do not rehydrate. The identification of vermiculite using these criteria can be illustrated by considering the X-ray diffractograms shown in Figure 54. The strong diffraction peak near 17 Å for the Ca-saturated glycerol and ethylene

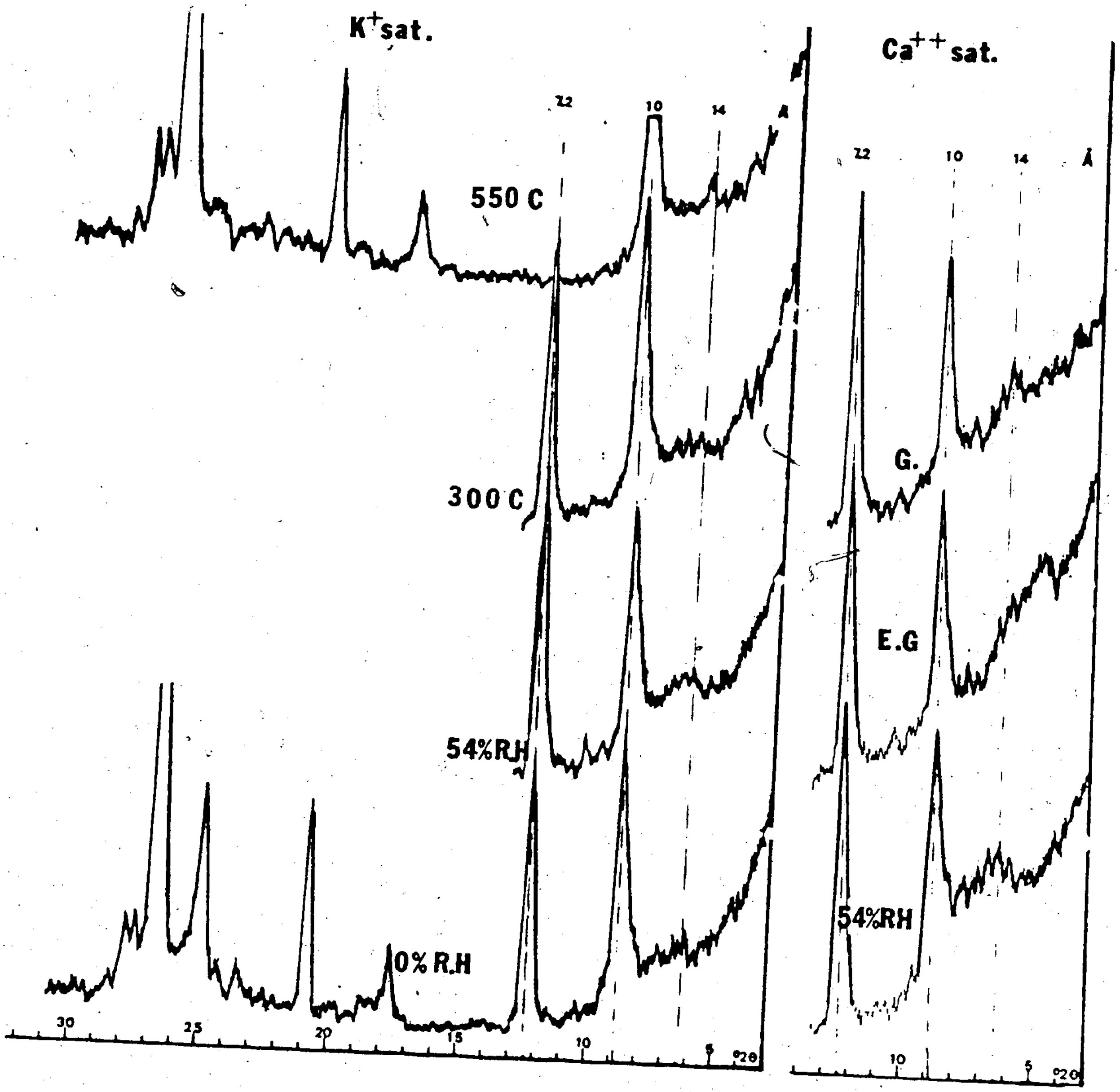


Figure 52. X-ray diffraction patterns of total clay separated from the Ae horizon of an Orthic Gray Luvisol (site 1)

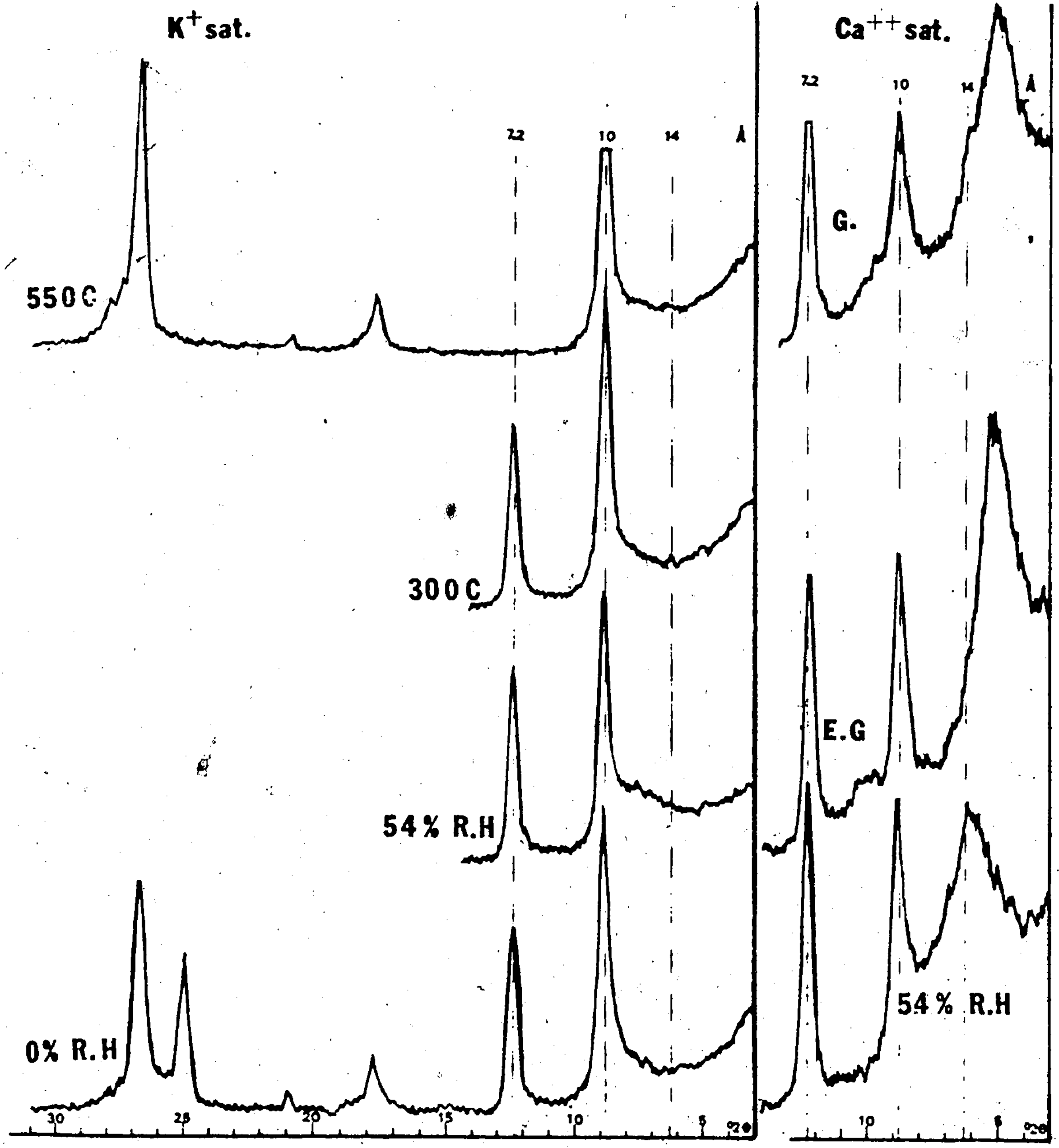


Figure 53. X-ray diffraction patterns of total clay separated from the Bt2 horizon of an Orthic Gray Luvisol (site 1)........................

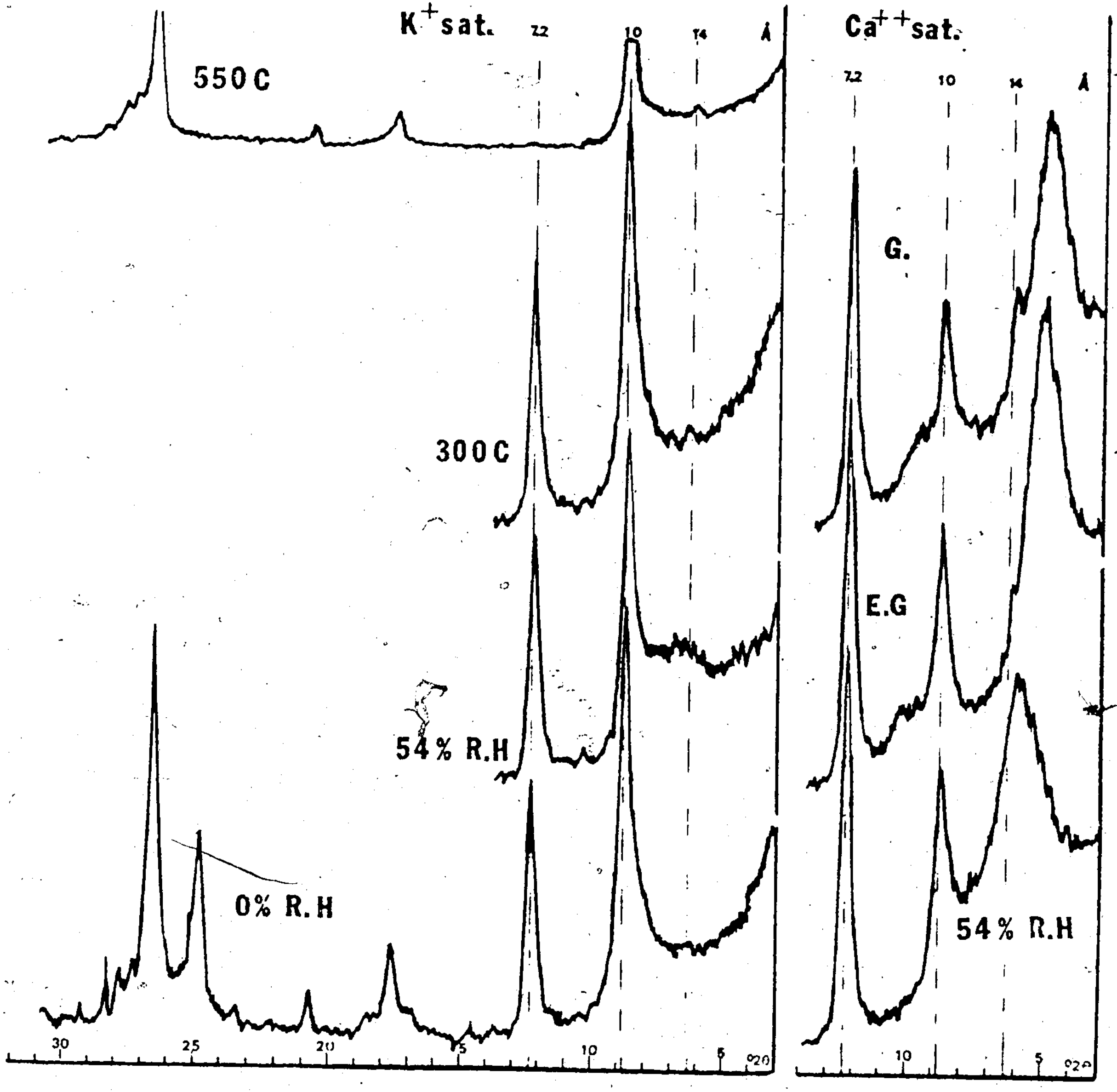


Figure 54. X-ray diffraction patterns of total clay separated from the Cca horizon of an Orthic Gray Luvisol (site 1)

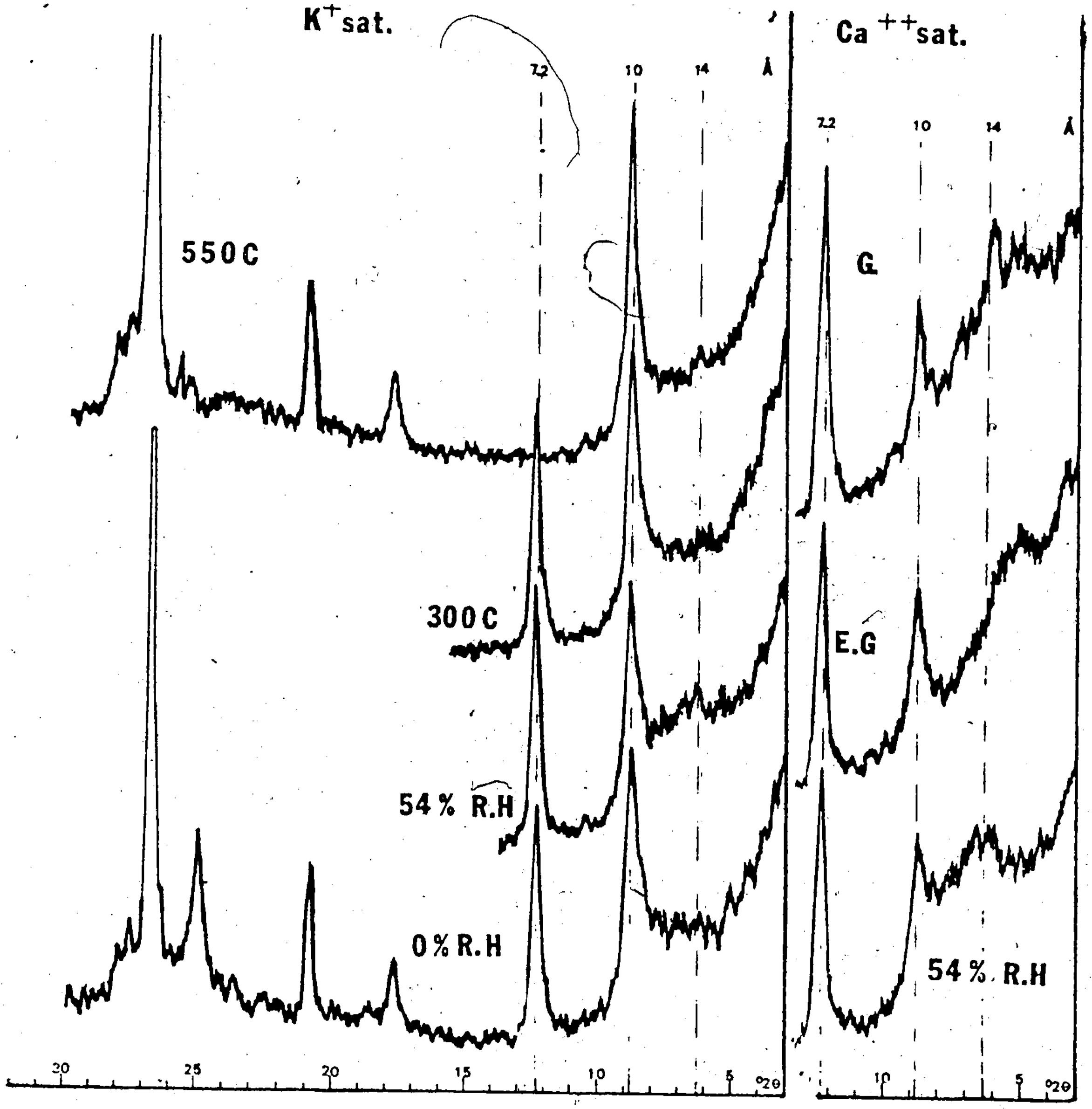


Figure 55. X-ray diffraction patterns of total clay separated from the Ahe horizon of a Dark Gray Luvisol (site 2)

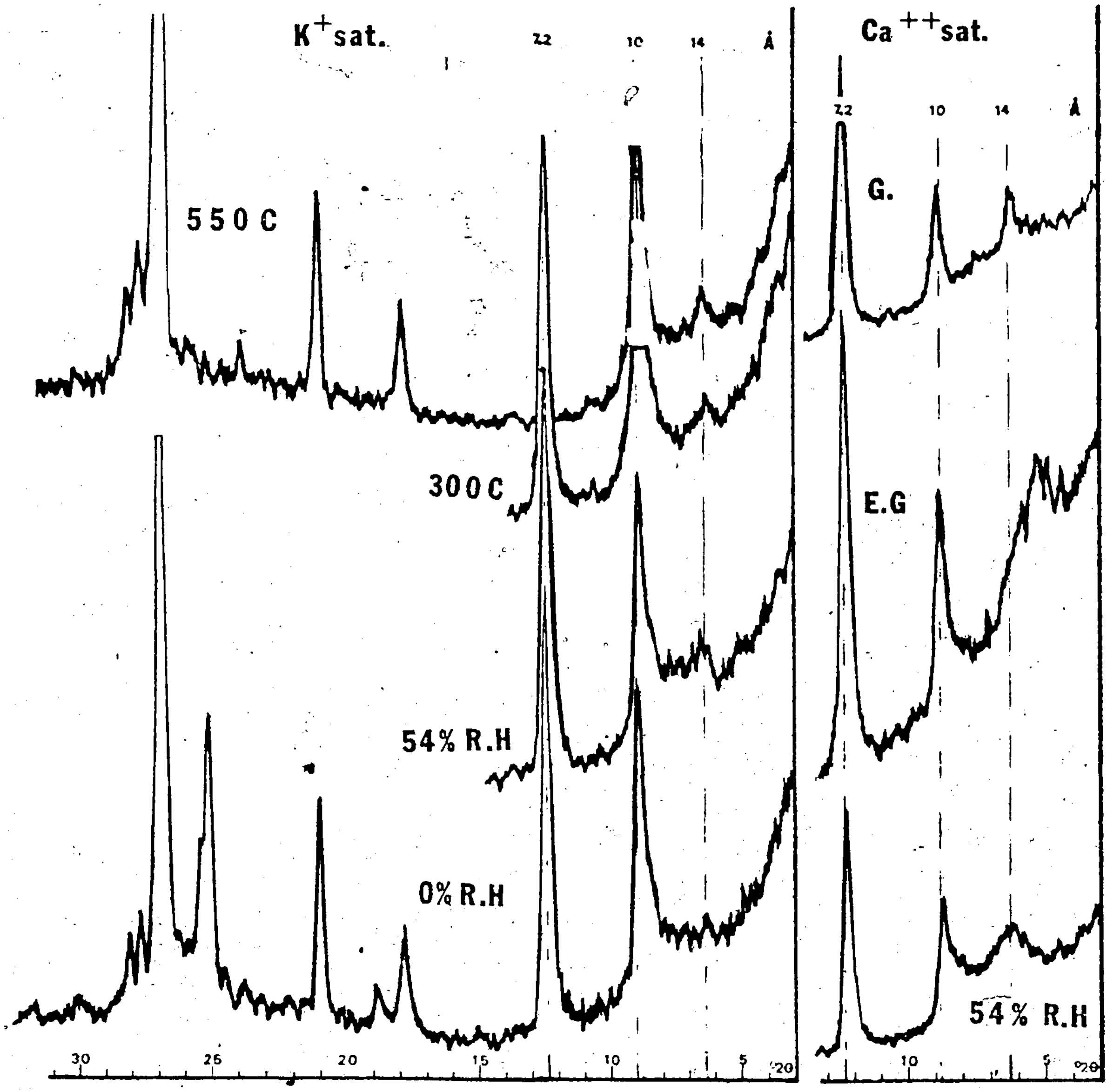


Figure 56. X-ray diffraction patterns of total clay separated from the Ae horizon of a Dark Gray Luvisol (site 2)...............................

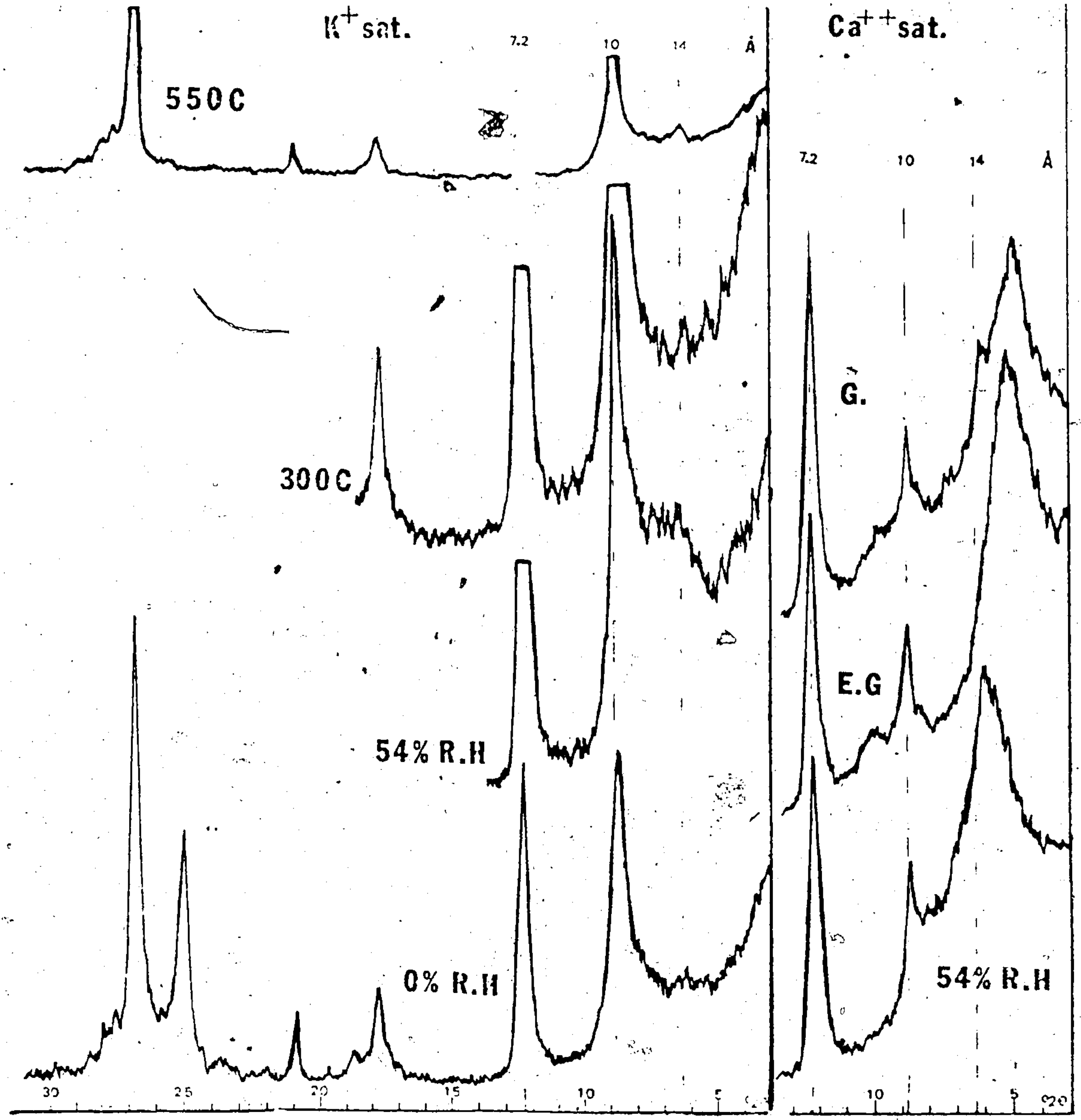


Figure 57. X-ray diffraction patterns of total clay separated from the Bt horizon of a Dark Gray Luvisol (site 2)..............................

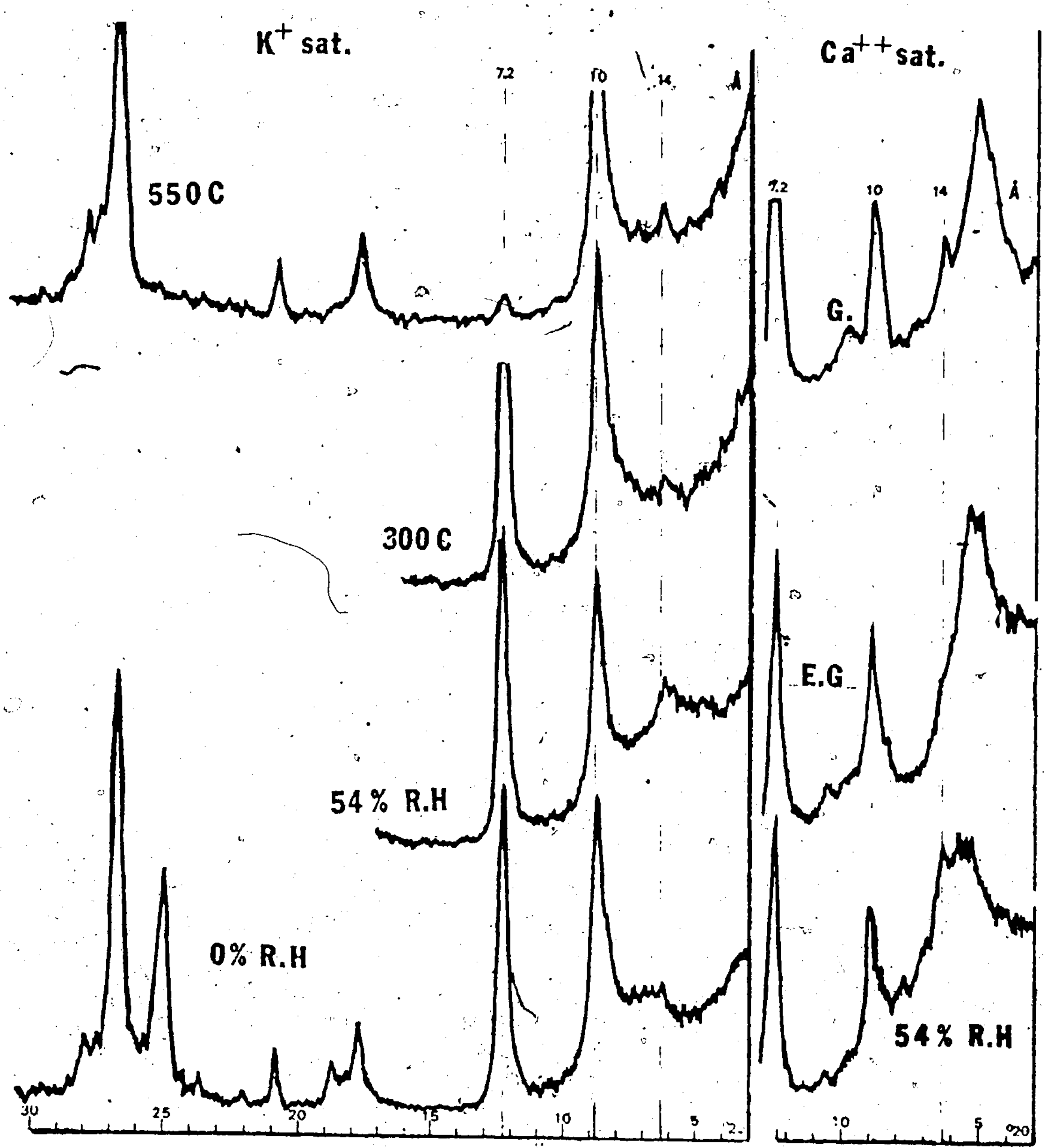


Figure 58. X-ray diffraction patterns of total clay separated from the Cca2 horizon of a Dark Gray Luvisol (site 2)........................

glycol solvated samples confirm the presence of smectite. The 14 Å peak occurring as a shoulder for the glycerol treated sample suggests vermiculite may be present; chlorite could also account for the 14 Å peak. For the K-saturated specimen heated to 105°C and held at 0% relative humidity during the X-ray analysis, all "expanded peaks" collapse to 10 Å and no 14 Å peak is evident which suggests chlorite is not present in any substantial amounts. Thus, the 14 Å for the Ca-saturated, glycerol solvated sample is probably due to soil vermiculite rather than to chlorite. This is confirmed by noting the degree of rehydration. For the K-saturated specimen (previously dessicated by heating at 105°C) equilibrated at 54% relative humidity, the intensity of the 14 Å is low and indicates that at least some of the expanded components observed in the Ca-saturated specimens are not rehydrating. Under these controlled conditions, the failure of rehydration can only be attributed to the presence of vermiculite. A slight variation of this line of reasoning is to compare the relative intensities of the 10 A:14-15 Å peaks of both K- and Ca-saturated specimens equilibrated at 54% relative humidity. Using the diffractograms displayed in Figure 54, this ratio is about 1 for the Ca-saturated specimen and about 6 for the K-saturated specimens indicating rehydration to the Ca-saturated status has not taken place. The only criticism of this later approach is that one is using the relative intensity of a broad peak (the 14 Å peak for the K-saturated specimen) in the comparison. To alleviate this possible criticism, one could simply compare the relative intensities of the 10 Å:7.2 Å peaks for Ca- and K-saturated samples. Applying this approach to the example under discussion (Figure 54), the intensity ratio would be 0.5 for the Ca-saturated specimen and 1.3 for the K-saturated specimen. Obviously, the increased intensity of the 10 Å peak relative to the 7 Å peak with K-saturation is due to the collapse and the inability of a clay mineral species to rehydrate. Vermiculite is the only species accountable for such behavior. Emphasis on accurate identification of vermiculite by both X-ray and chemical methods is included in this thesis because many soil clay mineral investigators in past studies have either ignored considering the presence of vermiculite or have based their findings of vermiculite on rather tenuous methods.

X-ray diffraction data, such as the diffractograms shown in figures 52 to 58, was used primarily for qualitative purposes to support the data and trends obtained by physical and chemical methods of quantitative clay mineralogy. Diffraction analysis showed that all clay separates contained smectite, mica and kaolinite as dominant clay mineral species. Vermiculite was usually detected in all samples by X-ray methods; quartz was also present while chlorite and chloritic intergrades were observed in some samples.

Smectite appeared depleted in clay separates from Ae horizons and high in relative abundance in B and C horizon samples as indicated by the intensity of 17 Å peaks for solvated specimens (Figures 52 to 54) which agrees with the profile distribution of smectite as determined by surface area and cation exchange capacity methods (Tables 15 to 20). The nature of smectite species was investigated using the Greene-Kelly test designed to differentiate montmorillonite from beidellite (the two common smectite species in soils) on the basis of diffraction behavior of Li-saturated and heated (220°C) specimens. A 001 spacing for specimens solvated with glycerol of about 9.5 Å indicates montmorillonite while a spacing of about 17.7 Å indicates beidellite (Brown, 1961). X-ray analysis revealed both montmonillonite and beidellite constitute the smectite phase (Figure 59) with beidellite concentrated in the fine clay (<.2 micrometer) fraction. Montmorillonite appeared to be more dominant than beidellite on the basis of X-ray diffraction characteristics. These findings concerning the presence of both montmorillonite and beidellite agree with the few studies previously reporting the occurrence of some beidellite in bedrock formations (Byrne and Farvolden, 1959), tills (Pawluk and Bayrock, 1969) and soils (Greene-Kelly, 1959) of Alberta. Since beidellite is present in the clay fractions and since it can fix potassium like vermiculite, the method used to determine quantity of vermiculite may include some contribution of beidellite.

Table 15. Physical, elemental and mineralogical analysis of clay.

Site : 1
Classification: Orthic Gray Luvisol

A)Physical and elemental analysis

Horizon	C.E.C. CAEC	KEC	S.A.	Na2O	K2O	CAO	MgO	FE2O3
	me/100g		m2/g			%		
Ae	28.2	19.4	147	0.95	3.03	0.24	1.75	5.78
AB	40.1	26.7	323	0.53	3.07	0.11	2.40	7.73
Bt1	50.7	35.9	370	0.59	3.23	0.11	2.50	8.81
Bt2	51.1	35.2	360	0.67	3.12	0.17	2.49	8.79
BCk	48.0	32.6	349	0.63	3.26	0.80	2.55	9.11
Cca	41.4	32.7	328	0.71	3.11	2.59	2.97	8.72

B)Mineralogical analysis.

	O.M.	CARB.	MICA	VERM.	SMEC. C.E.C.	SMEC. S.A.	KAOL.	QTZ.	CHL.	FLDS. PLG.	KF
					%						
Ae	ND	0.0	27	6	17	11	27	18	1	2	2
AB	ND	0.0	29	9	25	30	25	8	1	1	1
Bt1	ND	0.0	31	10	34	35	18	5			
Bt2	ND	0.0	29	10	33	33	21	5			
BCk	ND	1.1	31	10	31	32	21	4	1	1	1
Cca	ND	4.2	29	6	31	34	28	3	1	1	1

Abbreviations:

C.E.C. = Cation exchange capacity . S.A. = Surface area
O.M. = Organic matter . CARB. = Carbonates
VERM. = Vermiculite . SMEC. = Smectites
KAOL. = Kaolinite . QTZ. = Quartz
CHL. = Chlorite
KEC = Potassium exchange capacity, CAEC = Ca exchange capacity
FLDS. = Feldspars (PLG. for Plagioclases and KF for K-feldspars)

Table 16. Physical,elemental and mineralogical analysis of clay.

Site : 2
Classification: Dark Gray Luvisol.

A)Physical and elemental analysis.

Horizon	C.E.C. CAEC	 KEC	S.A.	Na2O	K2O	CAO	MgO	FE2O3
	me/100g		m2/g			%		
Ahe	44.7	25.7	200	0.54	2.92	0.16	2.00	7.73
Ae	24.2	20.2	152	0.64	3.06	0.14	1.96	6.33
Bt	46.9	38.2	386	0.44	2.68	0.12	2.19	9.40
Cca1	43.2	32.4	304	0.63	2.90	4.68	3.58	8.59
Cca2	43.1	32.1	312	0.77	2.93	3.79	3.47	8.52

B)Mineralogical analysis.

	O.M.	CARB.	MICA	VERM.	SMEC. C.E.C.	SMEC. S.A.	KAOL.	QTZ.	CHL.	FELDS. PLG.	 KF
						%					
Ahe	ND	0.0	24	12	24	12	29	19	>0.5	1	3
Ae	ND	0.0	28	3	19	15	35	11	1	1	2
Bt	ND	0.0	26	6	36	41	26	4	1	>0.5	>0.5
Cca1	ND	8.0	26	7	31	30	27	5	1	1	2
Cca2	ND	6.5	27	7	30	31	26	5	2	1	2

Abbreviations:

C.E.C. = Cation exchange capacity . S.A. = Surface area
O.M. = Organic matter . CARB. = Carbonates
VERM. = Vermiculite . SMEC. = Smectites
KAOL. = Kaolinite . QTZ. = Quartz
CHL. = Chlorite
KEC = Potassium exchange capacity. CAEC = Ca exchange capacity
FLDS. = Feldspars (PLG. for Plagioclases and KF for K-feldspars)

Table 17. Physical,elemental and mineralogical analysis of clay

Site : 3
Classification: Dark Gray luvisol.

A)Physical and elemental analysis

Horizon	C.E.C. CAEC me/100g	 KEC	S.A. m2/g	Na2O	K2O	CAO %	MgO	FE2O3
Ahe	47.2	28.9	187	0.59	3.72	0.17	1.20	7.01
Ae	31.3	22.6	187	0.52	3.07	0.12	1.91	6.98
AB	50.5	37.1	403	0.28	2.89	0.07	2.22	9.39
Bt1	55.2	38.0	418	0.36	2.85	0.07	2.33	9.04
Bt2	52.0	36.9	398	0.79	2.82	0.16	2.51	9.40
BC	43.1	33.4	363	0.57	2.88	1.06	2.69	8.92
Ck	43.2	31.2	324	0.72	3.03	4.53	3.15	8.80

B)Mineralogical analysis

	O.M.	CARB.	MICA	VERM.	SMEC. C.E.C.	SMEC. S.A. %	KAOL.	QTZ.	CHL.	FELDS. PLG.	 KF
Ahe	21.8	0.0	32	12	27	10	23	18	>0.5	2	3
Ae	ND	0.0	28	6	21	17	31	10	1	1	2
AB	ND	0.0	27	9	35	40	22	5	>0.5	>0.5	1
Bt1	ND	0.0	27	11	36	40	20	4	1	1	1
Bt2	ND	0.0	26	10	35	39	24	3	1	1	1
BCk	ND	1.9	27	6	22	38	27	5	1	1	1
Cca	ND	7.8	29	8	29	32	28	4	1	1	1

Abbreviations:

C.E.C. = Cation exchange capacity
O.M. = Organic matter
VERM. = Vermiculte
KAOL. = Kaolinite
CHL. = Chlorite
S.A. = Surface area
CARB. = Carbonates
SMEC. = Smectites
QTZ. = Quartz
KEC = Potassium exchange capacity, CAEC = Ca exchange capacity
FLDS. = Feldspars (PLG. for Plagioclases and KF for K-feldspars)

Table 18. Physical, elemental and mineralogical analysis of clay.

Site : 4
Classification: Orthic Gray Luvisol.

A)Physical and elemental analysis

Horizon	C.E.C.		S.A.	Na20	K20	CAO	MgO	FE203
	CAEC	KEC						
	me/100g		m2/g			%		
Ae	30.7	20.3	159	0.47	2.75	0.06	1.70	5.70
AB	55.7	35.8	379	0.36	2.50	0.02	2.20	7.78
Bt1	58.3	39.8	426	0.37	3.17	0.05	4.52	8.85
Bt2	58.1	38.8	421	0.38	3.41	0.09	4.60	9.06
Ck	54.5	37.2	417	0.33	2.81	1.24	2.48	9.19

B)Mineralogical analysis.

	O.M.	CARB.	MICA	VERM.	SMEC. C.E.C.	SMEC. S.A.	KAOL.	QTZ.	CHL.	FLDS. PLG.	KF
						%					
Ae	ND	0.0	24	7	19	12	30	16	1	1	2
AB	ND	0.0	23	13	34	36	23	4	1	>0.5	1
Bt1	ND	0.0	30	12	37	40	17	2	>0.5	>0.5	1
Bt2	ND	0.0	32	13	36	39	14	3	>0.5	>0.5	1
Ck	ND	---	27	11	35	40	21	4	1	>0.5	1

Abbreviations:

C.E.C. = Cation exchange capacity , S.A. = Surface area
O.M. = Organic matter , CARB.= Carbonates
VERM. = Vermiculite , SMEC.= Smectites
KAOL. = Kaolinite , QTZ. = Quartz
CHL. = Chlorite
KEC = Potassium exchange capacity, CAEC = Ca exchange capacity
FLDS. = Feldspars (PLG. for Plagioclases and KF for K-feldspars)

Table 19. Physical, elemental and mineralogical analysis of clay.

Site : 5
Classification: Eluviated Black Chernozemic.

A)Physical and elemental

Horizon	C.E.C. CAEC	 KEC	S.A.	Na2O	K2O	CAO	MgO	FE2O3
	me/100g		m2/g			%		
Ah	56.4	34.1	232	0.46	2.80	0.16	2.07	8.05
Ae	40.5	28.0	258	0.46	2.69	0.15	2.19	7.17
Bt1	51.0	38.7	417	0.55	2.66	0.17	2.41	8.82
Bt2	47.1	36.5	390	0.62	2.84	0.18	2.60	8.81
BCk	43.9	35.5	373	0.62	2.94	3.17	3.22	8.29
Cca	41.2	33.9	356	0.83	2.85	7.81	3.32	8.53

B)Mineralogical analysis.

	O.M.	CARB.	MICA	VERM.	SMEC. C.E.C.	SMEC. S.A.	KAOL.	QTZ.	CHL.	FLDS. PLG.	 KF
						%					
Ah	31	0.0	24	15	32	13	30	13	1	2	2
Ae	ND	0.0	25	8	26	23	30	8	1	1	1
Bt1	ND	0.0	26	8	37	43	25	4	>0.5	0	0
Bt2	ND	0.0	28	7	34	41	27	3	1	0	0
BCk	ND	6.3	28	6	34	40	24	5	1	1	1
Cca	ND	13.6	27	5	32	39	28	5	1	1	1

Abbreviations:

C.E.C. = Cation exchange capacity , S.A. = Surface area
O.M. = Organic matter , CARB. = Carbonates
VERM. = Vermiculite , SMEC. = Smectites
KAOL. = Kaolinite , QTZ. = Quartz
CHL. = Chlorite
KEC = Potassium exchange capacity, CAEC = Ca exchange capacity
FLDS. = Feldspars (PLG. for Plagioclases and KF for K-feldspars)

Table 20. Physical,Elemental and mineralogical analysis of clay.

Site : 6
Classification: Orthic Gray Luvisol.

A)Physical and elemental analysis

Horizon	C.E.C.		S.A.	Na2O	K2O	CAO	MgO	FE2O3
	CAEC	KEC						
	me/100g		m2/g			%		
Ae	35.5	24.7	258	0.58	2.28	0.08	1.72	5.31
AB	52.2	37.5	406	0.27	2.05	0.05	2.19	9.07
Bt	58.4	42.8	423	0.36	2.01	0.09	2.33	9.71
BCk	57.4	38.9	403	0.37	2.09	1.64	2.62	9.78
Cca	49.0	35.2	348	0.56	2.36	9.38	1.74	8.98

B)Mineralogical analysis.

	O.M.	CARB.	MICA	VERM.	SMEC. C.E.C.	SMEC. S.A.	KAOL.	QTZ.	CHL.	FLDS.	
										PLG.	KF
						%					
Ae	ND	0.0	22	7	23	24	35	12	>0.5	>0.5	>0.5
AB	ND	0.0	21	10	35	40	29	4	1	0	0
Bt	ND	0.0	20	10	40	42	25	4	1	0	0
BCk	ND	2.9	21	12	37	38	24	5	1	0	0
Cca	ND	16.7	24	9	33	33	26	7	1	0	0

Abbreviations:

C.E.C. = Cation exchange capacity , S.A. = Surface area
O.M. = Organic matter , CARB. = Carbonates
VERM. = Vermiculte , SMEC. = Smectites
KAOL. = Kaolinite , QTZ. = Quartz
CHL. = Chlorite
KEC = Potassium exchange capacity, CAEC = Ca exchange capacity
FLDS. = Feldspars (PLG. for Plagioclases and KF for K-feldspars)

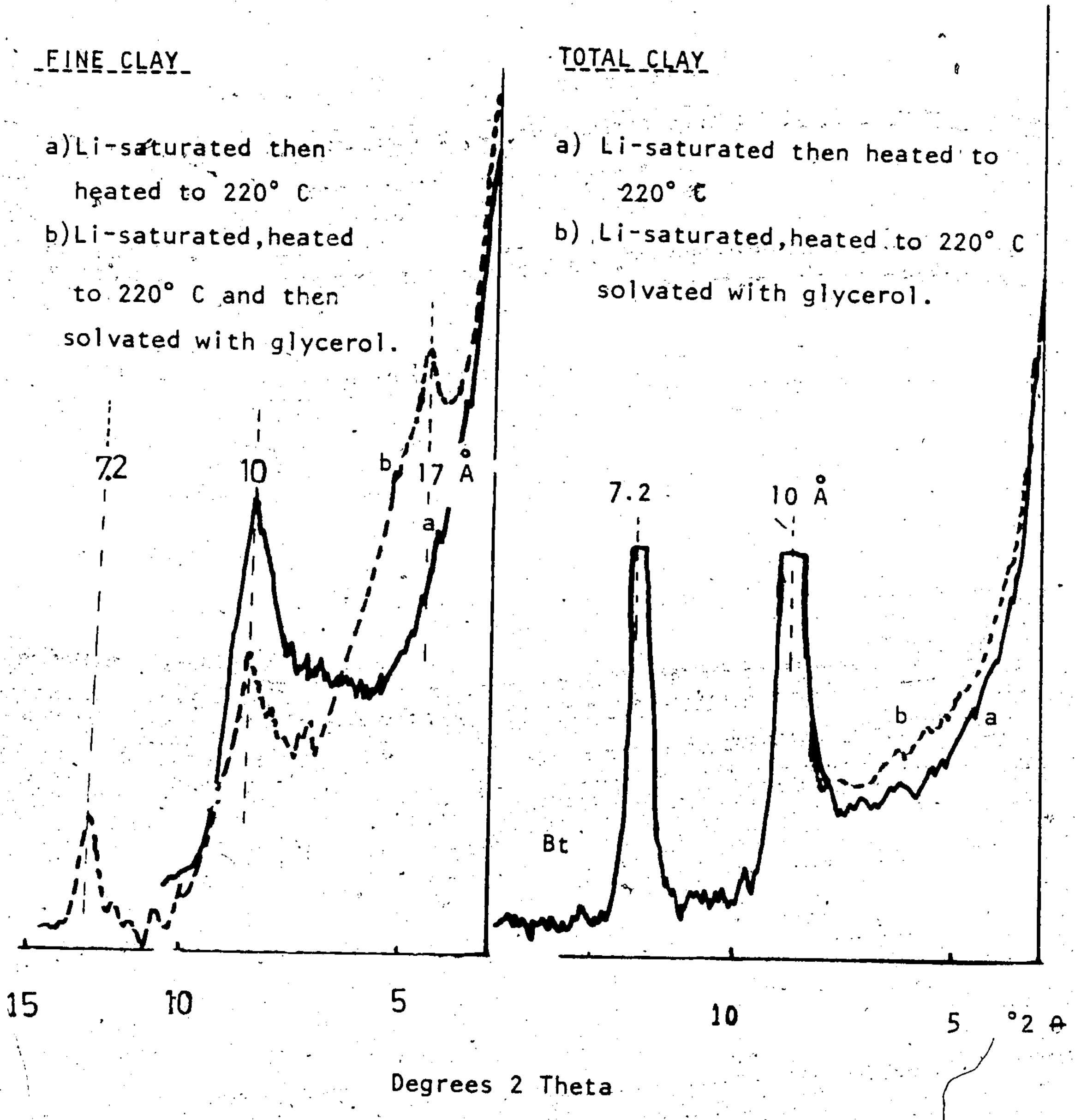
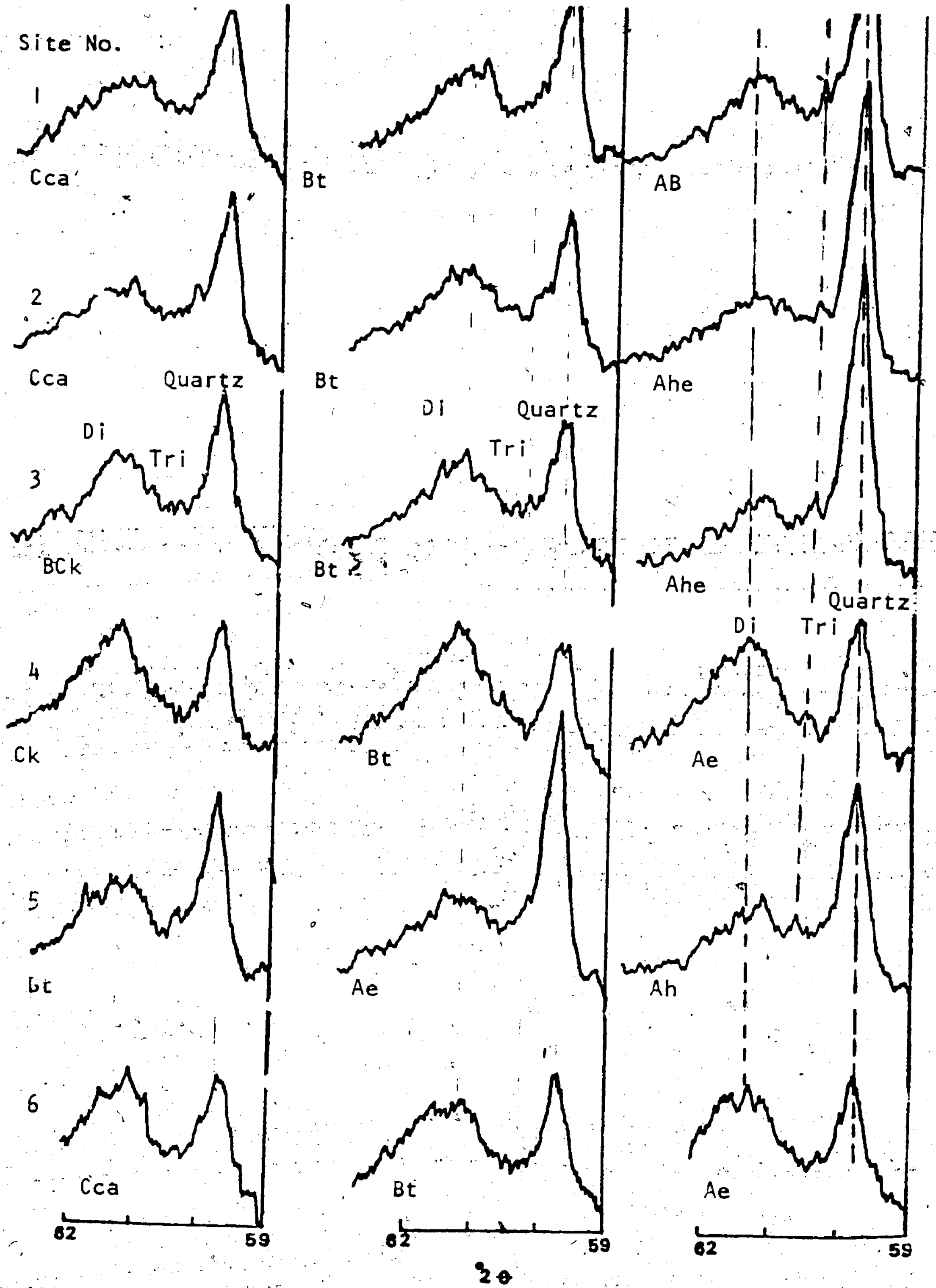


Figure 59. Representative X-ray diffraction patterns for Li-saturated clay a) for total clay and b) for fine clay (Greene-Kelley test)........

X-ray diffraction patterns also suggested kaolinite was in highest proportions in leached A horizons; its relative content seemed to diminish in B and C horizons as evidenced by the change in the intensity of the 7.2 Å peak. Diffraction peaks due to quartz and feldspar were always the most intense for clay separates from A horizons and indicated they (especially quartz) contributed substantially to the composition of the clay fraction.

Both chlorite and chloritic intergrades were detected in about one-half of the samples. They occurred in very small quantities as evidenced by the low intensity of their diagnostic X-ray diffraction peaks. The presence of a 14 Å peak and/or higher order reflections for the 550°C heated specimens is diagnostic for chlorites while chloritic intergrades display broad peaks or shoulders between 10 Å and 14 Å for K-saturated specimens analyzed under 0% relative humidity. The shoulder on the high angle side of the 10 Å peak for the K-saturated, 0% relative humidity treatment in Figure 58 is indicative of the presence of chloritic intergrades. Since the 001 peak intensities of chlorite were always very low or not detectable, the intensity of the 7.2 Å was thought to be essentially related to the quantity of kaolinite present in the samples.

X-ray diffraction patterns for the 59 to 62° 2θ region suggests the 2:1 type phyllosilicates are predominantly of a dioctahedral nature (Figure 60). Since these patterns were



<u>Abbreviations</u>: Di = Dioctahedral minerals, Tri = Trioctahedral

Figure 60. X-ray diffraction patterns of randomly oriented clay separated from the major genetic horizons..........................

obtained by analysis of previously heated and randomly oriented powder clay specimens, contributions of dioctahedral kaolinite will not appear on diffractograms since the kaolinite structure is destroyed during the heating of specimens. The dominance of dioctahedral versus trioctahedral phyllosilicates is reflected by the higher intensity of the dioctahedral 060 spacings located near 61.3° 2θ than the characteristic 060 trioctahedral spacing in the region of 60 to 61° 2θ. The minor trioctahedral component is probably mica (biotite) since trioctahedral smectites are of very rare occurrence in soils (Borchardt, 1977). The bulk of the clay separates, excluding kaolinite, would consist of dioctahedral smectites (montmorillonite and biedellite) and dioctahedral mica (muscovite). The prevalence of dioctahedral mica rather than trioctahedral species is also suggested by the relative intensity of 001:002 mica reflections for oriented clay specimens. This ratio of peak intensities was about 3:1 for all clay separates using the diffractograms for Ca-saturated, ethylene glycol solvated specimens. Dioctahedral species typically display relatively strong 002 peak intensities while trioctahedral varieties have strong 001 peak intensities and weak or absent 002 peaks so that the ratio of intensities of the 001:002 peaks are about 10:1 for trioctahedral mica and about 3:1 for dioctahedral varieties (Brown, 1961). These results strongly suggest muscovite is the dominant mica mineral species in the clay-sized separates.

Quantitative clay mineral data, based largely on physical and chemical methods of clay analyses, are shown in Tables 15 to 20 for the 6 soils of this study. In general, the clay mineral essemblages are similar for the 6 sites with parent materials containing 30 to 35% smectite, 24 to 29% mica, 21 to 28% kaolinite, 5 to 11% vermiculite, 3 to 7% quartz, up to 3% feldspars and 0 to 2% chlorite. Amounts of quartz, feldspars and chlorite may not be accurate since their adundances were determined using peak intensities from X-ray diffractograms and the empirically derived intensity conversion factors of Bayliss et al. (1970). The amount of smectite reported in this study is generally lower than the amounts usually described for parent materials in Alberta. The difference is probably due to inclusion of vermiculite with the smectite component in previous studies - this writer knows of no attempts where previous investigators working with Alberta soils have quantified with accuracy the vermiculite component.

Quantitative differences in clay mineral assemblages are more pronounced within a given soil rather than among soils. Content of smectite, in all soils, is highest in B horizons and selectively depleted from A horizons. The profile distribution of smectite closely resembles the profile distribution of fine clay where the horizon with the highest content of smectite is usually the same horizon with the highest proportion of fine clay (Table 21). Kaolinite appears to be negatively enriched in A horizons (Tables 15 to 20). Profile distribution of mica and vermiculite do not display obvious consistent patterns. Profile distribution of mica is more or less uniform with depth. Ae horizons often appear selectively depleted of mica relative to C horizons. Since smectite is obviously selectively removed from the clay fraction of Ae horizons, it is surprising that the proportion of mica is not higher because of negative enrichment. It seems that some mica must have weathered to account for a lack of a relative increase of mica in Ae horizons as compared to C horizons. The weathering of mica could account for the slightly elevated levels of vermiculite in A and B horizons. Profile distribution of clay-sized primary minerals (quartz, K-feldspar and plagioclase) illustrates their enrichment in Ae, Ahe, Ah and some AB horizons which could result through negative enrichment when smectite is preferentially translocated downward or through positive enrichment by physical shattering of coarser-sized (sand and silt) particles to clay-sized material as has been suggested by St. Arnaud and Mortland (1963). The later possibility is quite valid since the content of quartz in clay separates from surface horizons is about 2 to 6 times higher than amounts in clay from C horizons. Negative enrichment alone could not account for such a large relative increase in quartz. Enrichment of clay-sized quartz in surface soil horizons has been noted by

Table 21. Content of smectite and fine clay in the total clay (<2 um) separate.

Soil	Horizon	% of total clay	
		Fine clay	Smectite[1]
Orthic Gray Luvisol site (1)	Ae	24	17
	AB	44	25
	Bt1	63	34
	Bt2	57	33
	BCk	52	31
	Cca	49	31
Dark Gray Luvisol (site 2)	Ahe	31	12
	Ae	32	19
	Bt	73	36
	Cca1	64	31
	Cca2	65	30
Dark Gray Luvisol (site 3)	Ahe	20	10
	Ae	31	21
	AB	62	35
	Bt1	58	36
	Bt2	61	35
	BCk	62	32
	Cca	58	29
Orthic Gray Luvisol (site 4)	Ae	24	19
	AB	56	34
	Bt1	61	37
	Bt2	58	36
	Ck	56	35
Eluviated Black Chernozemic (site 5)	Ah	30	13
	Ae	47	26
	Bt1	60	37
	Bt2	58	34
	BCk	57	34
	Cca	61	32
Orthic Gray Luvisol (site 6)	Ae	33	23
	AB	60	35
	Bt	63	40
	BCk	59	37
	Cca	64	33

other investigators (Huang and Lee, 1969; McKeague et al., 1972).

f. Fine clay fraction

Fine clay fractions (<0.2 micrometer) were characterized by X-ray diffraction methods only; as a consequence, only qualitative and semi-quantitative information was obtained for this size separate. Smectite, mica and kaolinite appear to be the sole crystalline constituents of the fine clay fractions from all soils. On the basis of diffraction peak intensities, smectite apparently dominates the fine clay fraction in all soil horizons (Figure 61). Diffraction peaks for the micaceous mineral at 10 Å for the Ca-saturated specimens tend to be broad and are indicative of some depotassification. Peak broadening is most conspicuous for clays from surface horizons suggesting weathering in the biologically active zone has resulted in some degradation of fine clay-sized mica. The relative proportion of smectite appears to increase with profile depth as suggested by the progressively higher intensity of the 17 Å peak proceeding from the A to the B to the C horizons (Figure 61). Content of X-ray amorphous materials, as indicated by the intensity of the sloping background (compared to the intensity of the 10 Å peak) from 3 to 8° 2θ for the K-saturated specimens appears highest in A horizon clay separates and diminishes with profile depth. Vermiculite was typically absent from fine clay fractions which means its relative content in the

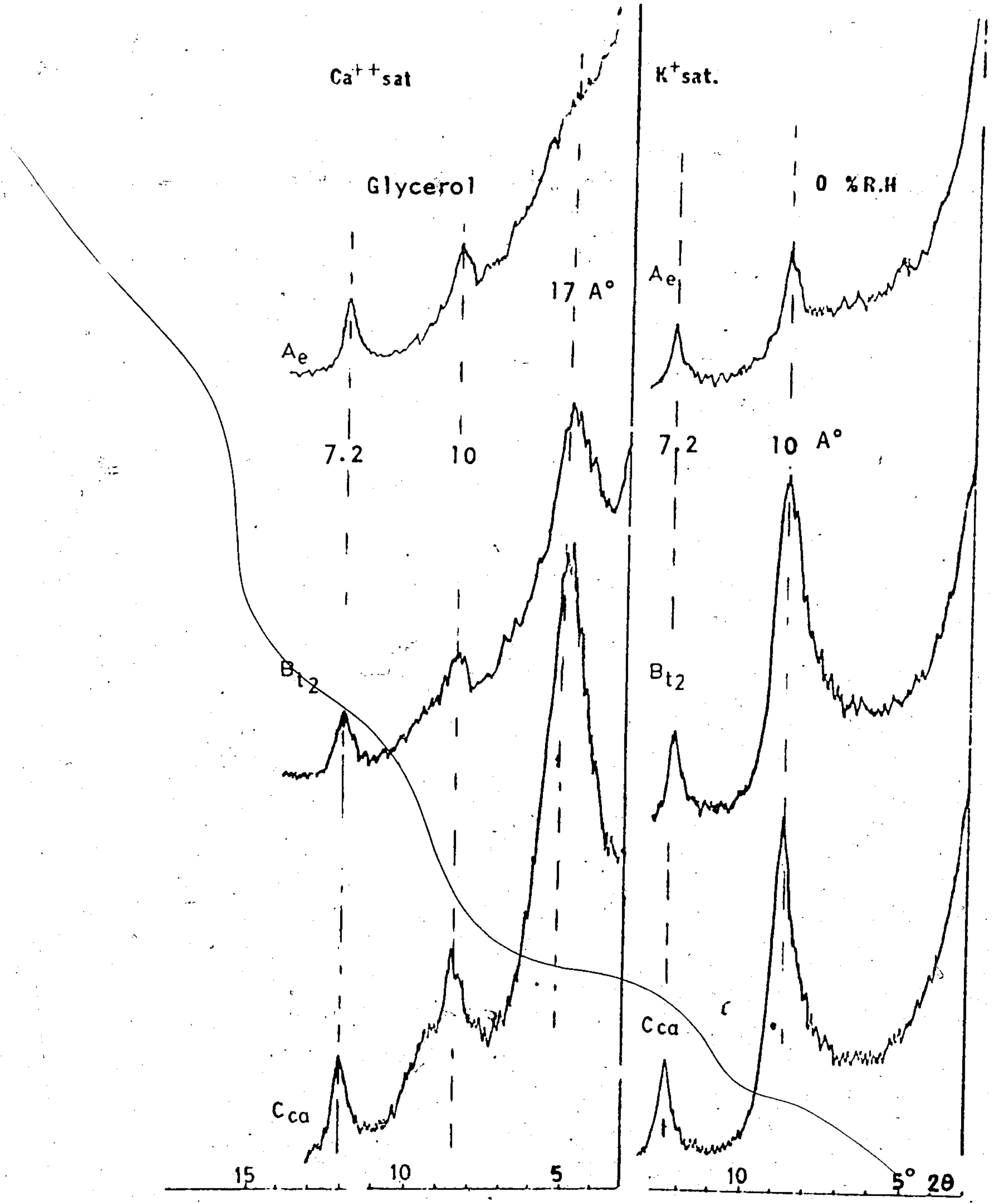


Figure 61. X-ray diffractograms of fine clay (<0.2 µm) from the A, B and C horizons of the Orthic Gray Luvisol at site 1........................

coarse clay fraction (2 to 0.2 micrometer) would be much higher than the amount determined for the total clay fraction (Table 15 to 20). It seems vermiculite is either unable to persist weathering when it occurs as fine clay-sized particles or its formation involves the development of relatively large sized crystals as would be the case when trioctahedral mica experiences depotassification.

A summary of all mineral constituents in the six soils of this study is presented in the last section of this thesis.

C. Parent Material Uniformity

The establishment of parent material uniformity in pedological studies is a necessary step in the evaluation of pedogenic changes within the soil solum. When the parent material of a soil is established to be relatively uniform, the changes in the upper solum can then be attributed to pedogenic processes.

Glacial till, the parent material for the soils of this study, is a complex and variable sediment (Flint, 1971 and Goldthwait, 1971). The lateral and vertical variations in physical characteristics of glacial till deposits have been recognized by many researchers in the field of Geology, Civil Engineering and Soil Science; the heterogeneous nature of till is related to differences in type and structure of the source bedrock, the initial and englacial transport action of the ice, the local incorporation of materials from the bedrock and the mode of deposition of these materials from the ice sheet (Scott, 1976).

In a study of the characteristics of Alberta tills, Pawluk and Bayrock (1969) pointed out the inherent and variable characteristics of the till and demonstrated that they largely reflect the nature of the underlying bedrock. Twardy (1969), in his study of tills in west-central Alberta, pointed out that the Cooking Lake till is different in texture from Breton and Hubalta tills despite the fact that all three tills occur in close proximity and were deposited by the Continental ice sheet. Differences in mode of deposition was thought to be the reason for the textural variability. Gravenor and Bayrock (1965) earlier recognized the regional variability in Alberta tills but they indicated in some localities tills have a remarkably uniform texture. Particle size analysis of fifty till samples from the Wainwright area in eastern Alberta revealed that 75 per cent were of silty clay loam texture while the remainder were loam to clay loam (Gravenor and Bayrock, 1965). Poor sorting and lack of stratification are widely recognized as important characteristics of tills but as Sudom and St. Arnaud (1971) pointed out, some tills are foliated or even bedded. Water action may cause sorting and washing of fine materials whereas eolian deposition may contribute fine particles to the supraglacial drift.

On the basis of the previous discussion, it is appropriate to consider and evaluate the possibility that the parent materials of the soils of this study are heterogeneous and contain lithologic discontinuities.

The following [illegible]ction deals with the assessment of parent material uniformity for the six study sites using the methods employed by pedologists in previous studies. The depth function of a certain parameter (e.g. quartz to feldspar ratio) could indicate the presence of a lithological discontinuity if the function displays a certain degree of departure from what would be acceptable for a uniform material. Unfortunately, there is no general agreement as to what degree of variation defines a discontinuity (Evans, 1978). Drees and Wilding (1973) in a study of elemental variation within a sampling unit in tills, loess and outwash suggested that if a lithological discontinuity is present between two horizons there should be a significant difference between these two horizons which exceeds the random variations (e.g. lateral variations within each horizon) by the value of t for a given probability level. Their study suggests that a relative standard deviation in the Ti:Zr ratio between two horizons must exceed 28% to indicate a significant lithological break between two horizons in a profile developed on glacial till. A relative standard deviation in excess of 100% in the Ti:Zr ratio for two horizons was required to confirm lithologic discontinuities in soils developed on sandstones, siltstones and cherty limestone (Chapman and Horn, 1968). Deviation from the mean of 4.8 to 21% for Zr:Sr ratios and 8.1 to 17.6% for quartz: illite ratios were considered to indicate parent material uniformity in soils developed from scree deposits of Lower Paleozoic mudstones (Evans and Adams, 1975). From the wide differences in the level of variation accepted for a uniform material it would seem that the nature of the deposit, the nature of the parameter used (e.g. quartz:feldspar ratio or Ti:Zr ratio, etc.) and the accuracy of the determination for that parameter are important factors in determining the acceptable level of variation for a uniform material.

In this study, horizons within each profile were considered as individual uniform units (thus, assuming a lateral and vertical uniformity within each soil horizon) and the possibility of discontinuities between these units (soil horizons in each profile) was evaluated on the basis of deviation from the mean (relative standard deviation in per cent) for a given parameter as compared to generally accepted deviation for a uniform glacial till published in the literature.

Depositional stratification was not evident in the field, and observed changes in color, texture and structure within each profile were attributed to pedological processes rather than to the presence of any lithological discontinuity. Various methods were used to further asses the apparent uniformity seen in the field. The parameters used include particle size data, content and distribution of resistant minerals (e.g. quartz and feldspars) and the depth function of the content of tri-acid residue of the soil.

a. Particle Size Distribution

Data on particle size distribution has been employed by many researchers as a basis for interpreting lithological discontinuities in soils derived from loess or glacial till (St. Arnaud and Whiteside, 1963; McKeague and Brydon, 1970; Miller et al., 1971; Sudom and St. Arnaud, 1971; Sneddon et al., 1972; Evans and Adams, 1975; Price et al., 1975; Rutledge et al., 1975).

Since particle size distribution within the soil solum is affected by clay translocation, it is necessary to use parameter values calculated on a non-clay basis even though such calculations may bias the data. Silt distribution on a clay-free basis within profiles (Figures 62 and 63) indicates: (1) no apparent change at site 6, (2) 14% less silt in the Ae horizon than the average silt content of the remaining horizons for site 1 (relative deviation of 37%), (3) 10 to 12% more silt in the Ahe horizon of site 2, in the Ah and Ahe horizons of site 3, and in the Ah horizon of site 5 than the average content in the remaining horizons of respective profiles, and (4) at site 4, silt content of Ae, AB and Bt1 horizons are about 12% higher than the average content in the Bt2 and Cca horizons (a relative deviation of

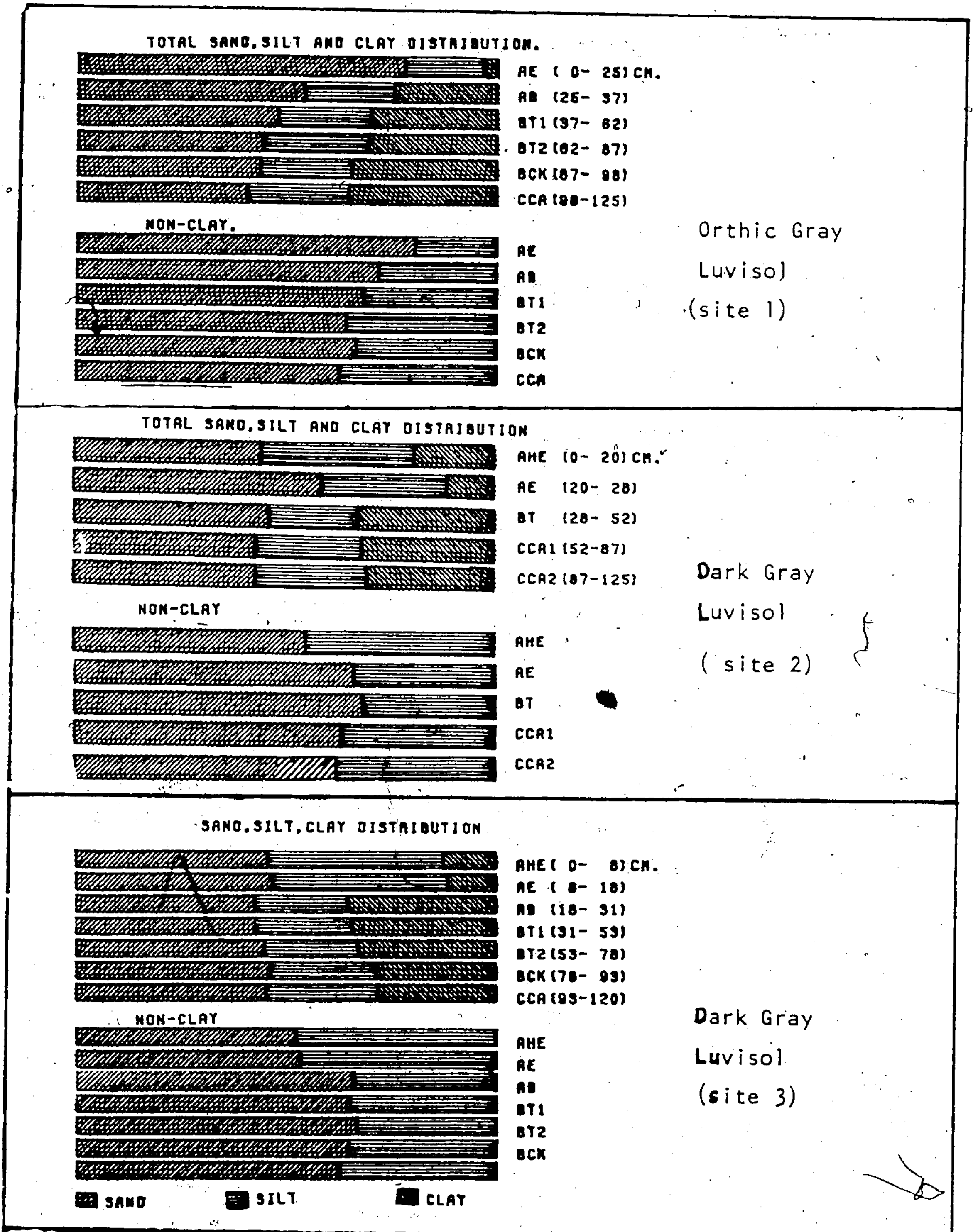


Figure 62. Total and non-clay particle size distribution for the major genetic horizons (sites 1 to 3).......................

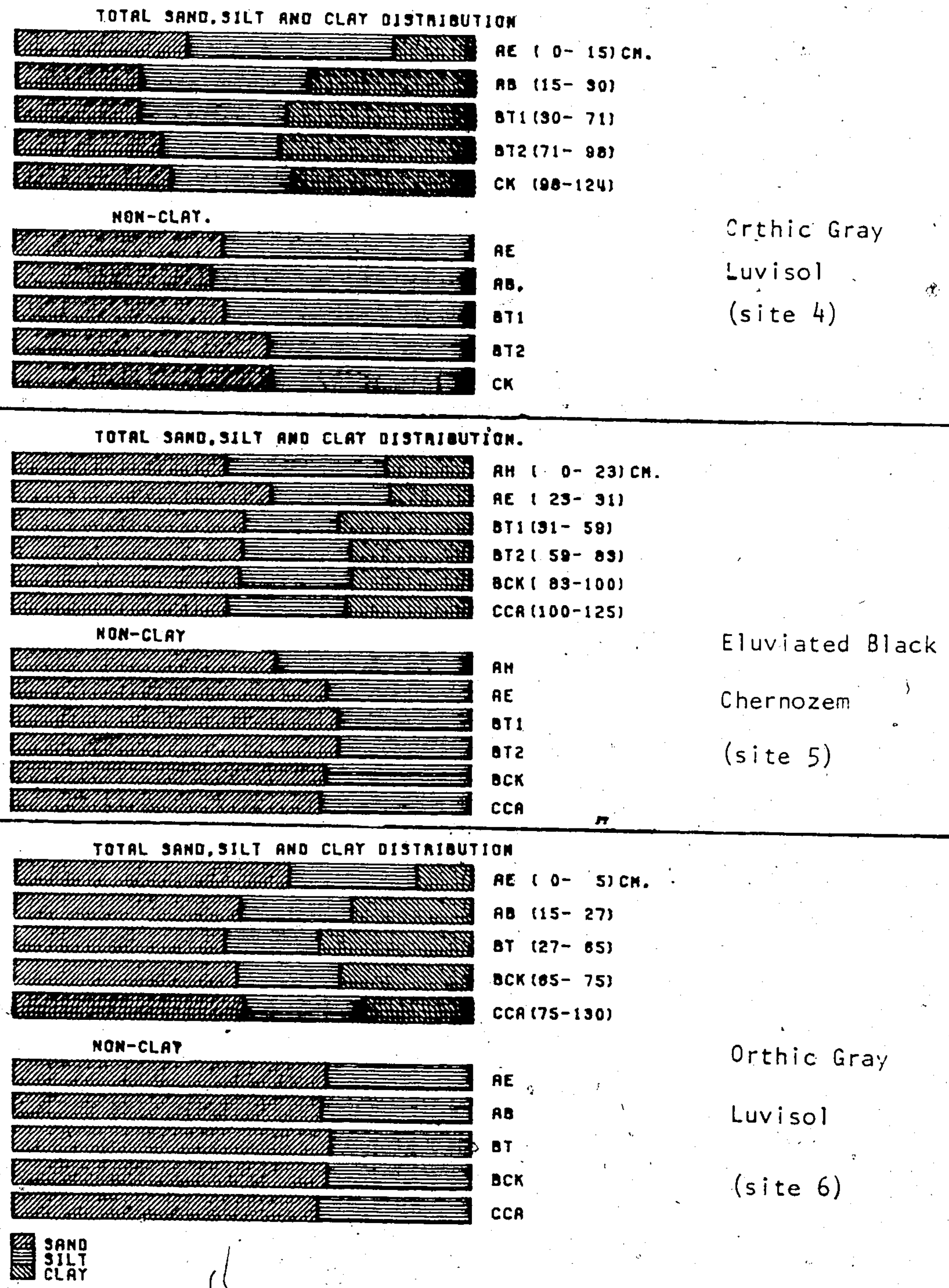


Figure 63. Total and non-clay particle size distribution for the major genetic horizons (sites 4 to 6)......................

17%). The results of silt distribution in the non-clay fraction indicate either there is a parent material dicontinuity at the above mentioned breaks in the content of silt or that processes associated with pedogenesis may explain the observed results. The higher content of silt in the upper 18 to 23 cm at sites 2, 3, 4 and 5 when compared to the underlying horizons may have been caused by simple additions of silt-sized materials of eolian origin and/or due to physical disintegration of particles coarser than silt to silt-sized material (a pedogenic cause) (St. Arnaud and Whiteside, 1963). Mechanical breakdown of gravel and sand-sized materials has been attributed mainly to frost action in surface horizons of Luvisolic and Chernozemic soils which were periodically moisture-saturated. In a freeze-thaw experiment (200 cycles) conducted by St. Arnaud and Whiteside (1963) on water-saturated sands and gravels (>0.5 mm in size) separated from C horizons of Luvisolic soils, 1.2 to 3.8% of fine particles were produced by frost action in their laboratory experiments. Distribution of silts (on a clay-free basis) in at least sites 2, 3 and 5 indicate the lowest amounts occur in B horizons while amounts in C horizons are closer to the mean content of silt in the apparently enriched surface horizons. This suggests that some silt in Bt horizons may have undergone breakdown to clay-sized material. Hence, a pedogenic rather than a lithologic cause may account for the observed differences in silt content between horizons in profiles 2, 3 and 5. The uniform silt distribution at site 6 is indicative of parent material uniformity. Although pedogenic processes (e.g. physical weathering) may have contributed to the high silt content (on clay-free basis) in the surface Ae horizon at site 4, it would not explain the higher content in AB and Bt horizons when compared to the lower Bt and Ck horizons. However, variation from the mean (the relative standard deviation of content of silt in the entire profile) in the content of silt is about 17% which is not large enough to indicate a lithologic discontinuity. The differences in silt content throughout the profile may simply reflect minor depositional variation in till at site 4. At site 1, the content of silt (on clay-free basis) in the Ae horizon is 14% less than the average content of silt in the remaining horizons. This distribution is peculiar since the other profiles do not display a similar trend and may be a result of the presence of sand pockets which were observed at the time of sampling. Additions of eolian sand or removal of fine particles by either wind or water action from the surface of the till after deposition could also be a contributing factor. The content of silt in the Ae horizon of this profile is 19.5% (calculated on a clay-free basis) while the average content of silt in all the underlying horizons is 33.3 ± 3.6%. This amounts to a relative standard deviation of 37% based on the content of silt in the Ae horizon and the average content of silt in the remainder of the profile. The implication of this relatively high value

(37%) for site 1, will be discussed at the end of this section.

Ratios for fine to total sand, coarse and medium to fine sand and fine sand to total silt were also used as criteria for establishing parent material uniformity as has been done by others (Barshad, 1964; Evans and Adams, 1975). The ratios indicate small variations within each profile (1 to 23% deviation from the means) except for sites 1 and 4 where the variations in some ratios are as high as 39 and 62% (Tables 22 and 23). These values indicate lithologic breaks do not occur in profiles 2, 3, 5 and 6. The coarse and medium sand to fine sand ratios at site 4 deviate (relative standard deviation) by 60% from the mean and suggest a possible lithologic break between the AB and Bt horizons. At site 1, the fine sand to total silt ratio indicates some large variation present between the Ae horizon and the remainder of the profile (Table 22) indicative of a possible lithologic discontinuity.

The parent materials of profiles 2, 3, 5 and 6 also appear uniform on the basis of cumulative-frequency curves for particle size distribution for the non-clay fraction (Figure 64). Cumulative curves for the upper most horizons from sites 2, 3 and 5 show a slight difference from those of lower horizons in the size range 5 to 250 micrometer which may be related to pedological reasons (e.g. physical disintegration of coarse particles) and/or some eolian

Table 22. Physical characteristics of soils related to particle size distribution (sites 1 to 3).

	Particle size ratios		
Soil horizons	Fine sand to Total sand	C&M sand to Fine sand	Fine sand to Total silt
<u>Orthic Gray Luvisol (site 1)</u>			
Ae	0.87	0.14	3.60
AB	0.87	0.15	2.21
Bt1	0.86	0.17	1.85
Bt2	0.85	0.17	1.52
BCk	0.88	0.14	1.75
Cca	0.85	0.18	1.42
Mean ± S.D	0.86 ± 0.01	0.16 ± 0.02	2.06 ± 0.80
%deviation from the mean	1	13	39
<u>Dark Gray Luvisol (site 2)</u>			
Ahe	0.93	0.08	1.15
Ae	0.86	0.16	1.71
Bt	0.86	0.16	1.87
Cca1	0.89	0.12	1.55
Cca2	0.87	0.15	1.44
Mean ± S.D	0.88 ± 0.03	0.13 ± 0.03	1.54 ± 0.27
%deviation from the mean	3	23	18
<u>Dark Gray Luvisol (site 3)</u>			
Ahe	0.93	0.08	1.01
Ae	0.86	0.16	0.97
AB	0.87	0.15	1.67
Bt1	0.88	0.14	1.64
Bt2	0.87	0.15	1.74
BCk	0.89	0.12	1.63
Cca	0.89	0.12	1.50
Mean ± S.D	0.88 ± 0.02	0.13 ± 0.03	1.45 ± 0.32
%deviation from the mean	2	23	22

Table 23. Physical characteristics of soils related to particle size distribution (sites 4 to 6).

	Particle size ratios		
Soil horizons	Fine sand to Total sand	C&M sand to Fine sand	Fine sand to Total silt
Orthic Gray Luvisol (site 4)			
Ae	0.90	0.11	0.76
AB	0.91	0.10	0.69
Bt1	0.71	0.42	0.60
Bt2	0.74	0.36	0.92
Ck	0.68	0.47	0.87
Mean ± S.D	0.79 ± 0.11	0.29 ± 0.18	0.77 ± 0.13
%deviation from the mean	14	62	17
Eluviated Black Chernozemic (site 5)			
Ah	0.61	0.64	0.82
Ae	0.63	0.60	1.36
Bt1	0.63	0.59	1.55
Bt2	0.64	0.56	1.38
BCk	0.61	0.64	1.24
Cca	0.68	0.47	1.24
Mean ± S.D	0.63 ± 0.03	0.58 ± 0.06	1.27 ± 0.25
%deviation from the mean	4	10	20
Orthic Gray Luvisol (site 6)			
Ae	0.62	0.63	1.32
AB	0.62	0.60	1.28
Bt	0.64	0.58	1.43
BCk	0.65	0.55	1.40
Cca	0.63	0.60	1.24
Mean ± S.D	0.63 ± 0.01	0.59 ± 0.03	1.33 ± 0.08
%deviation from the mean	2	5	6

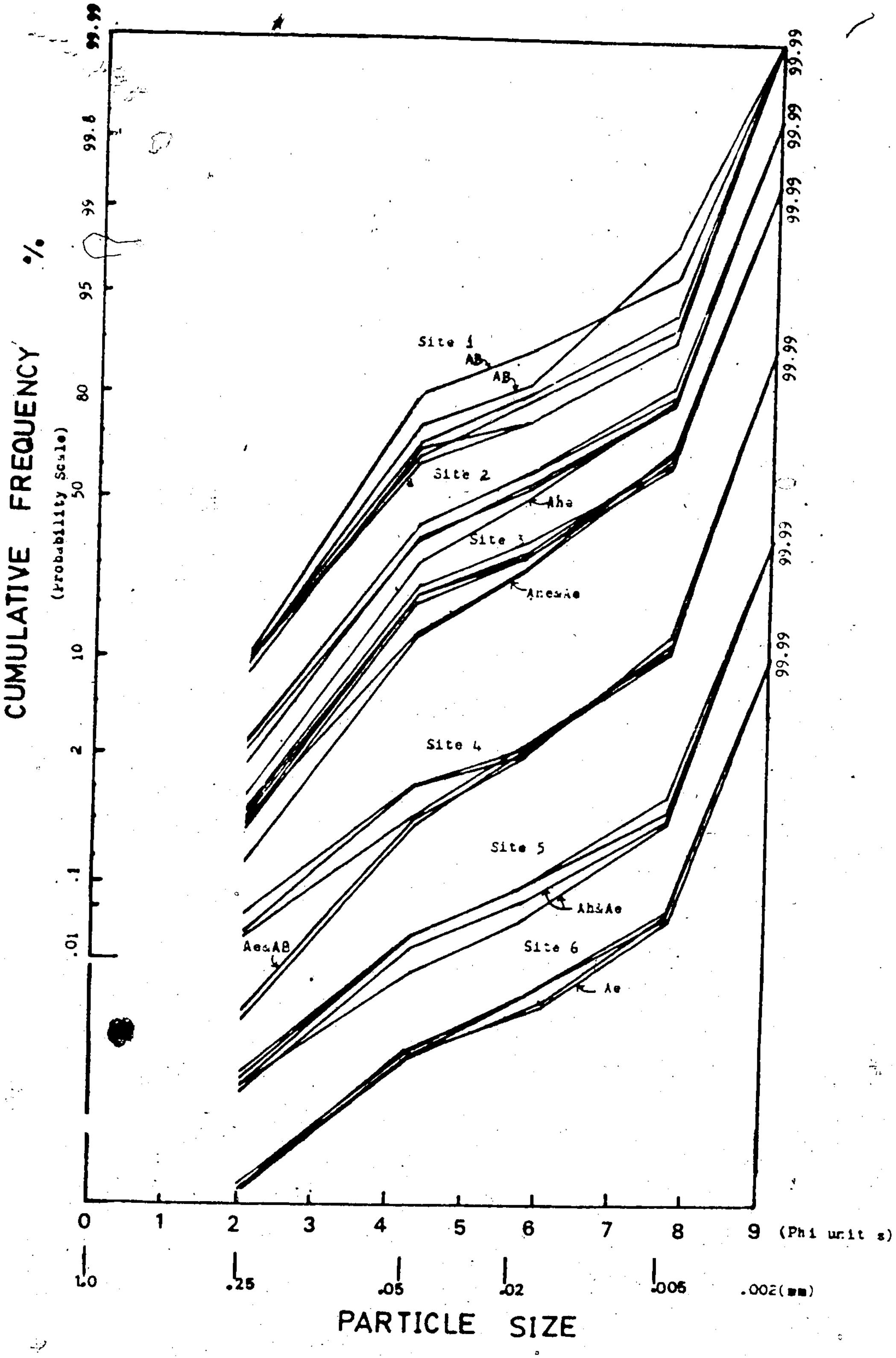


Figure 64. Cumulative-frequency curves for soil horizons for the six soils.....................

additions of particles in this size range into the surface. At site 1 and 4, cumulative-frequency curves for the Ae horizon (site 1) and the Ae and AB horizons (site 4) are unlike those for underlying horizons and suggests some modification of the materials at these two sites. Cumulative-frequency curves reflect sorting (the extent to which particle sizes are clustered about the mean grain size) and "distribution of the initial particle sizes of the parent material which the sorting process affected" (Friedman and Sanders, 1978). Several approximate statistical parameters are used to evaluate the frequency-distribution curves such as graphic mean, inclusive graphic standard deviation and graphic kurtosis. Lately, these parameters were used to indicate the occurrence of lithologic discontinuities (Stewart et al., 1977). Distribution of these parameters (mean-Mz, sorting-OI and kurtosis-KG) with depth confirm earlier suggestions of uniform parent materials for soils at sites 2, 3, 5 and 6 (Table 24). Variations of up to 0.46 phi units in the mean-Mz, 0.26 phi units in sorting -OI and 0.21 phi units in Kurtosis-KG were observed for soils at sites 2, 3, 5 and 6. The highest variations occur within sites 1 and 4 with variations of 0.73 to 0.86 phi units in the mean, 0.56 phi units in sorting and 0.72 to 0.11 phi units in Kurtosis. Variations of up to about 1.6 phi units for the mean and 1 phi unit for sorting and kurtosis were necessary to indicate the presence of two lithologic units of loess and andesitic

Table 24. Approximate grain size parameters (values in phi units).

	Statistical parameters*		
Horizon	MZ	OI	KG
O.G Luvisol (site 1)			
Ae	3.48	1.46	1.60
AB	3.90	1.66	1.24
Bt1	4.04	1.86	1.16
Bt2	4.18	1.91	1.07
BCk	4.34	2.01	0.88
Cca	4.33	2.02	0.91
D.G Luvisol (site 2)			
Ahe	4.53	1.76	0.93
Ae	4.09	1.84	1.11
Bt	4.07	1.86	1.14
Cca1	4.31	1.87	0.99
Cca2	4.27	1.89	1.00
D.G Luvisol (site 3)			
Ahe	4.60	1.74	0.91
Ae	4.45	1.85	0.92
AB	4.25	1.89	0.91
Bt1	4.23	1.85	0.98
Bt2	4.14	1.86	1.09
BCk	4.27	1.85	0.95
Cca	4.31	1.86	0.98
O.G Luvisol (site 4)			
Ae	4.84	1.85	0.81
AB	4.96	1.86	0.84
Bt1	4.58	2.28	0.90
Bt2	4.37	2.28	0.79
Ck	4.23	2.41	0.82
E.BLK. Chernozemic (site 5)			
Ah	3.95	2.51	0.89
Ae	3.55	2.33	1.04
Bt1	3.60	2.30	1.06
Bt2	3.64	2.25	1.08
BCk	3.62	2.49	0.98
Cca	3.86	2.31	1.04
O.G Luvisol (site 6)			
Ae	3.48	2.38	1.09
AB	3.52	2.32	1.03
Bt	3.55	2.32	1.08
BCk	3.69	2.37	1.00
Cca	3.64	2.33	1.02

* Parameters are: MZ=Mean; OI=Sorting; KG=Kurtosis

tephra (Stewart et al., 1977). In profile 1, variations in the grain size parameters with depth suggests uniformity in the profile if the Ae horizon is excluded. In the Ae horizon, the mean and sorting values are less while Kurtosis is higher when compared to underlying horizons. The decrease in the mean value is a result of an abundance of coarse particles while the decrease in the sorting value indicates improvement of sorting. Glacial till deposits are usually poorly sorted, thus improvement of sorting in the Ae horizon of site 1 may have occurred as a result of fine particles being stripped from the deposit by either wind and/or water action. In profile 4, the mean grain size of Ae and AB horizons is lower than that of the remainder of the profile and is indicative of an abundance of fine particles. The value for sorting(OI) are lower in A horizons and suggests some sorting of materials in Ae and AB horizons (additions of fine size materials and/or depositional variations between the upper and lower solum).

The cumulative-frequency curves for the non-clay fraction for the surface and C horizons of profiles 1 and 4 are compared, together with the curves of North American loess (Smith, 1942), lacustrine silts, glacial till and for a surficial deposit over till. The surficial deposit(over till) was thought to be a mixture of till and eolian silts of glacial origin (Arnold and Cline, 1961) (Figure 65). There is little or no similarity in size distribution between surface horizons of soils of this study and loess

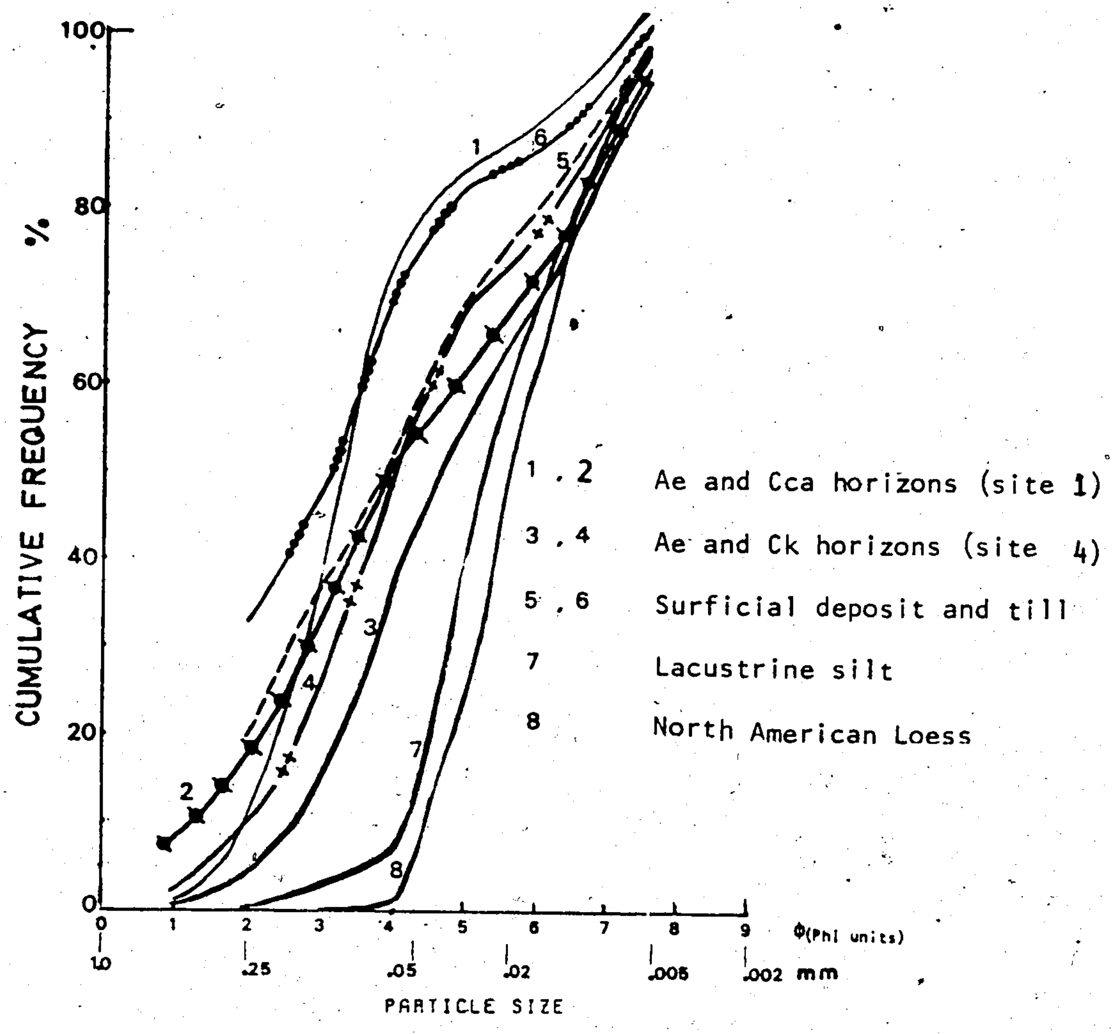


Figure 65. Cumulative-frequency curves for surface horizons and C horizons from sites 1 and 4, a surficial deposit (thought to approximate a 3:1 mixture of till and lacustrine silt, after Arnold and Cline, 1961) and North American loess (after Smith, 1942)............

material for particle sizes larger than 0.1 mm. The shape of the curve for the one Ae horizon (site 4) is similar to that of lacustrine silts and loess for the particle size range of less than 3 phi units (0.125 mm). Material in this size range is readily wind-borne. Glacial till, unlike lacustrine silts, contains particles in the range of 250 to 2000 micrometer and thus cumulative curves of mixtures of silts and glacial till will tend to be displaced upwards from the curve for silt alone (curve 7, Figure 65). On this basis, it appears that the surface horizon of profile 4 probably contains some eolian material deposited on the surface of the profile (compare the curve for the C horizon with that of the Ae horizon) (4 versus 3, Figure 65). The curve for surficial material (5) overlying till (6) (Arnold and Cline, 1961) shows the same downward shift from till as observed for the surface horizons of site 4. The cumulative-frequency curve for the Cca horizon from site 1 (2) is similar to that for the Ck horizon of of site 4 (4). However, the curve for the Ae horizon of site 1 (1) is displaced upwards unlike that of the Ae horizon of site 4 (4). This upward displacement suggests a different process at site 1 such as depletion of fine particles by wind and/or water action.

b. Resistant mineral distribution

Quartz has been used as a stable reference mineral by numerous researchers (Cann and Whiteside, 1955; St. Arnaud and Whiteside, 1963; Al-Janabi and Drew, 1967; Redmond and Whiteside, 1967; Smeck et al., 1968; Sudom and St. Arnaud, 1971; Evans, 1978). Its high resistance to weathering make it a suitable indicator mineral for studies of weathering and parent material uniformity. The relatively high percentage of quartz (in comparison to other minerals) is advantageous since errors introduced from determination and/or sampling will be relatively small. Other minerals such as K-feldspars could also be used to indicate parent material uniformity especially in relatively young soils where they are expected to display little or no weathering. Barshad (1964) suggested the ratio of resistant minerals is more reliable than percentages in evaluating the uniformity within soil profiles.

Amounts of both quartz and K-feldspars and their ratios for the 5 to 250 micrometer size range were calculated for the total non-clay fraction. The distribution of quartz within all profiles shows a small but gradual decrease with increasing profile depth for the coarse and medium silt; this trend is also apparent for quartz in the fine sand fraction (Tables 7 to 12, Appendix A). K-feldspars generally display a reverse trend. Variations (relative standard deviations in per cent) in contents of quartz, K-feldspars and their ratios within profiles and for various size fractions suggests materials are uniform in all sites except for site 4 which displays the highest deviation from the means in all variables (Table 25). Size distribution of quartz within individual profiles in the various non-clay fractions are quite similar for soils from sites 2, 3, 5 and

Table 25. Variations in content of quartz, K-feldspar, K-feldspar: quartz ratios and tri-acid residues within profiles.

Variable	Size	Deviation from the means within profiles* %					
		site 1	site 2	site 3	site 4	site 5	site 6
Quartz	fine sand	2	2	2	3	2	7
	C&M silt	3	4	4	5	4	2
	Nonclay**	6	4	3	11	3	1
K-feldspar	fine sand	5	5	4	4	7	7
	C&M silt	7	8	6	17	1	11
	Nonclay**	6	6	2	9	7	8
K-feldspar to Quartz Ratio	find sand	6	6	4	6	8	8
	C&M silt	9	12	10	20	3	13
	Nonclay**	9	7	2	14	6	9
Tri-acid residue	<2 mm	5	4	4	5	2	4

* relative standard deviation in per cent

** nonclay refers to the size fraction 2-0.002 mm.

6) as shown in the bar graphs and the ternary diagrams (Figures 66 and 67). The size distribution of quartz in soils from sites 1 and 4 is quite unlike distribution of quartz in all other profiles (Figure 66). At site 1, content of coarse and medium sand-sized quartz appears uniform throughout the profile (Figure 66). Coarse and medium silt-sized quartz decreases with profile depth while the reverse trend occurs for fine sand-sized quartz. An anomolous quartz distribution pattern is also evident for site 4 where content of coarse and medium sand-sized quartz is low in the Ae and AB horizons and much higher in the underlying horizons (Figure 66). Physical shattering of coarse and medium sand-sized quartz to fine sand-sized quartz may possibly account for the results. However, physical breakdown of coarse sand-sized quartz to fine sand-sized quartz appears to occur only in the surface horizons of two other profiles (sites 2 and 3). Physical disintegration of large particles to small particles is mainly attributed to frost action in surface soil horizons . Climatic conditions at all site locations are similar, thus the extent of physical disintegration should be similar at all six sites. Since not all profiles displayed evidence for physical disintegration of coarse sand-sized quartz to fine sand-sized quartz, the importance of this process is questionable. Minor eolian additions and depositional variations could account for the variations in quartz distribution with profile depth.

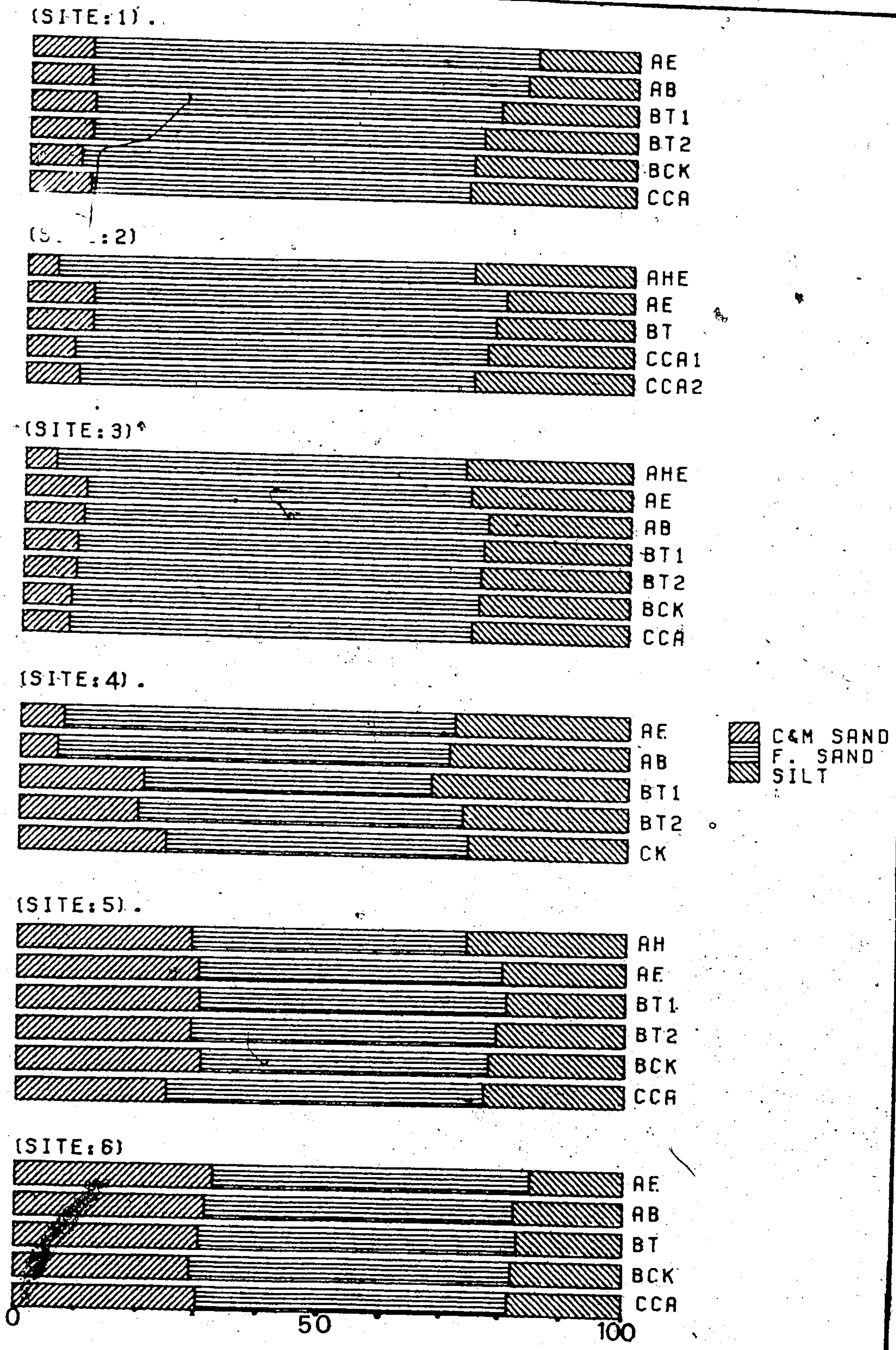


Figure 66. Bar graphs showing distribution of quartz in the various non-clay fractions................

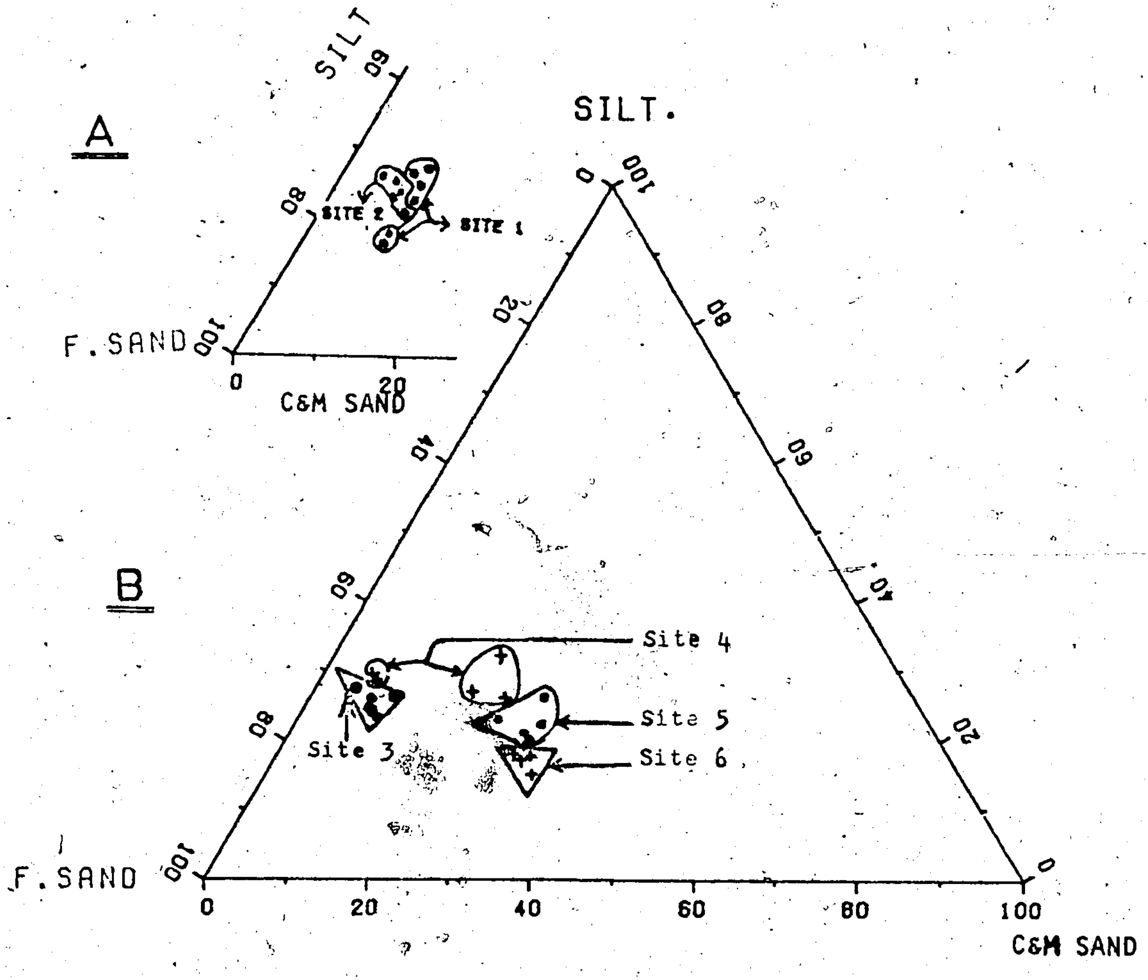


Figure 67. Ternary diagram showing size distribution of quartz in the non-clay fraction for soils from sites 1 and 2 (A) and sites 3 to 6 (B).

c. Tri-acid residue

Content of tri-acid (a mixture of HCl, HNO3 and H2SO4) residue was found to be a more suitable internal standard than contents of quartz, ZrO2 or TiO2 (Evans and Adams, 1975). Thus, distribution of tri-acid residue should be uniform for a uniform material.

The per cent deviation from the mean (relative standard deviation) of contents of acid residues for horizons within each soil are shown in Table 25. Profile distribution patterns of tri-acid residue were uniform within profiles. The rather small variations suggest parent materials of all profiles are uniform and do not contain discontinuities. The interpretation concerning parent material uniformity based on the variations of tri-acid residue does not agree entirely with the interpretations based on the other parameters used in this study which showed that some irregularities occur at site 1 and site 4. The reason for this difference in results could be due to the relatively insensitive nature of the tri-acid residue test which is based on the variation of all "resistant" material in the total <2 mm size fraction.

In conclusion, both textural and mineralogical information indicates lithologic breaks do not occur in profiles 2, 3, 5 and 6 while some modification of the parent material is evident for surface horizons at sites 1 and 4. The similarity in the qualitative nature of feldspars from upper and lower solum sections at sites 1 and 4 suggests there is no apparent mineralogical discontinuity. At site 1, some loss of fine particles, possibly through wind and/or water action has affected the particle size distribution where upper horizons (Ae and to a lesser extent AB horizons) are slightly depleted of fine particle sizes. At site 4, the data suggests upper horizons have received some eolian additions of fine materials. Small quantities of eolian material having mineralogical characteristics similar to the glacial till material of equivalent particle size, could also be present in surface horizons for soils at sites 2 and 5. Since mineralogical discontinuities were not detected in any of the soils, the general trends in mineral distribution within profiles could be interpreted qualitatively in terms of pedogenic effects. However, quantitative evaluation of pedogenic effects within soils from sites 1 and 4 are not valid due to non pedogenic modifications of their surface horizons.

D. Mineral Alteration During Pedogensis

Previous studies of soils from the Plains region of Western Canada suggest mineral weathering accompanying pedogenesis is not pronounced. Some of the characterization data obtained in this study revealed that minor mineral alteration has occurred in response to soil development. For example, content of heavy minerals within a given size fraction typically decreased proceeding down the soil profile. This section deals briefly with further quantitative evaluation of mineral weathering using bulk chemical analysis of total sand and silt fractions and mineral ratios.

In general, contents of elemental oxides in the sand and silt fractions are quite similar among the six soil profiles (Tables 26 and 27). Amounts of K2O and Na2O are one and half times greater in the silt as compared to the respective sand fractions while amounts of CaO are generally equal to or slightly more in the silt as compared to the sand fraction. These differences in elemental contents are a reflection of the differences in mineralogy of these two size fractions. Silt fractions contain one to one and half times as much potassium and sodium feldspar and two to three times as much heavy minerals as does the sand fraction. The largest relative difference in elemental oxide content within a given profile occurs for CaO and Na2O.

Weathering effects, as reflected by elemental distribution within a profile are more pronounced for silt fractions than for sands (Figure 68). This figure is quite typical of the elemental distribution patterns displayed by all six soils. Sand fractions did not display consistent patterns which may suggest that weathering effects in sands are not large and may have been masked by inherent elemental distribution in tills.

Table 26. Bulk Chemical analysis for total sands and silts (sites 1 to 3).

	Sand (2-0.02 mm)					Silt (0.02-0.002 mm)				
Horizon	K_2O	Na_2O	CaO	MgO	Fe_2O_3	K_2O	Na_2O	CaO	MgO	Fe_2O_3
	%									
Orthic Gray Luvisol (site 1)										
Ae	1.28	1.02	0.69	0.15	0.67	2.09	1.43	1.03	0.52	1.60
AB	1.27	1.01	0.72	0.19	0.72	2.27	1.40	1.04	0.53	1.66
Bt1	1.36	1.14	0.73	0.15	0.71	2.25	1.52	1.16	0.58	1.79
Bt2	1.33	1.15	0.80	0.15	0.76	2.24	1.55	1.20	0.60	1.88
BCk	1.37	1.05	0.72	0.12	0.61	2.48	1.62	1.36	0.61	1.92
Cca	1.43	1.10	0.83	0.16	0.70	2.38	1.56	1.31	0.59	2.00
Dark Gray Luvisol (site 2)										
Ahe	1.33	1.06	0.70	0.15	0.58	2.07	1.26	0.78	0.54	1.84
Ae	1.24	1.01	0.70	0.17	0.59	2.08	1.28	0.85	0.62	1.95
Bt	1.27	1.01	1.04	0.36	0.91	2.17	1.42	1.10	0.78	2.34
Cca1	1.50	1.18	1.05	0.31	0.71	2.06	1.49	1.24	0.56	1.58
Cca2	1.43	1.16	1.01	0.27	0.73	2.46	1.57	1.27	0.55	1.51
Dark Gray Luvisol (site 3)										
Ahe	1.31	1.20	0.85	0.19	0.85	2.01	1.30	0.86	0.51	1.57
Ae	1.48	1.17	0.84	0.24	0.89	2.06	1.29	0.87	0.55	1.73
AB	1.48	1.16	0.93	0.14	0.81	2.32	1.39	0.97	0.64	2.14
Bt1	1.36	1.21	0.78	0.13	0.58	2.23	1.46	1.03	0.64	1.94
Bt2	1.42	1.11	0.80	0.15	0.85	2.23	1.53	1.17	0.65	2.04
BCk	1.49	1.23	0.82	0.17	0.80	2.21	1.55	1.20	0.61	1.78
Cca	1.41	1.14	0.97	0.27	0.79	2.24	1.64	1.22	0.65	1.7

Table 27. Bulk Chemical analysis for total sands and silts (sites 4 to 6).

Horizon	Sand (2-0.02 mm) K20	Na20	CaO	MgO	Fe203	Silt (0.02-0.002 mm) K20	Na20	CaO	MgO	Fe203
	%									
	Orthic Gray Luvisol (site 4)									
Ae	1.11	0.68	0.44	0.14	1.50	1.81	1.09	0.43	0 52	2.14
AB	1.14	0.92	0.54	0.19	0.73	1.80	1.07	0.29	0.66	3.14
Bt1	1.05	0.78	0.46	0.07	0.62	1.88	1.15	0.46	0.71	2.72
Bt2	1.09	0.81	0.54	0.14	0.63	1.86	1.33	0.68	0.61	2.33
Ck	1.13	0.83	0.85	0.22	0.76	2.02	1.32	0.62	0.60	2.09
	Eluviated Black Chernozemic (site 5)									
Ah	1.18	0.90	0.63	0.16	0.63	1.89	2.09	1.24	0.81	1.71
Ae	1.04	0.74	0.53	0.15	0.58	1.93	1.27	0.86	0.50	1 64
Bt1	1.05	0.79	0.54	0.13	0.46	1.91	1.34	0.93	0.52	1.70
B[illegible]2	1.02	0.84	0.59	0.12	0.45	1.91	1.43	1.01	0.59	1.72
BCk	1.16	0.91	1.07	0.35	0.48	1.95	1.43	1.12	0.58	1.56
Cca	1.18	1.05	1.03	0.67	0.60	1.98	1.49	1.16	0.59	1.64
	Orthic Gray Luvisol (site 6)									
Ae	0.74	0.55	0.34	0.05	0.42	1.78	1.02	0.50	0.50	1.88
AB	0.83	0.54	0.31	0.07	0.56	1.75	1.19	0.65	0.56	1.86
Bt	0.68	0.43	0.30	0.07	0.45	1.72	1.15	0.75	0.57	1.84
BCk	0.68	0.47	0.39	0.12	0.44	1.69	1.16	0.75	0.53	1.78
Cca	0.70	0.60	0.64	0.19	0.58	1.67	1.25	0.77	0.53	1.74

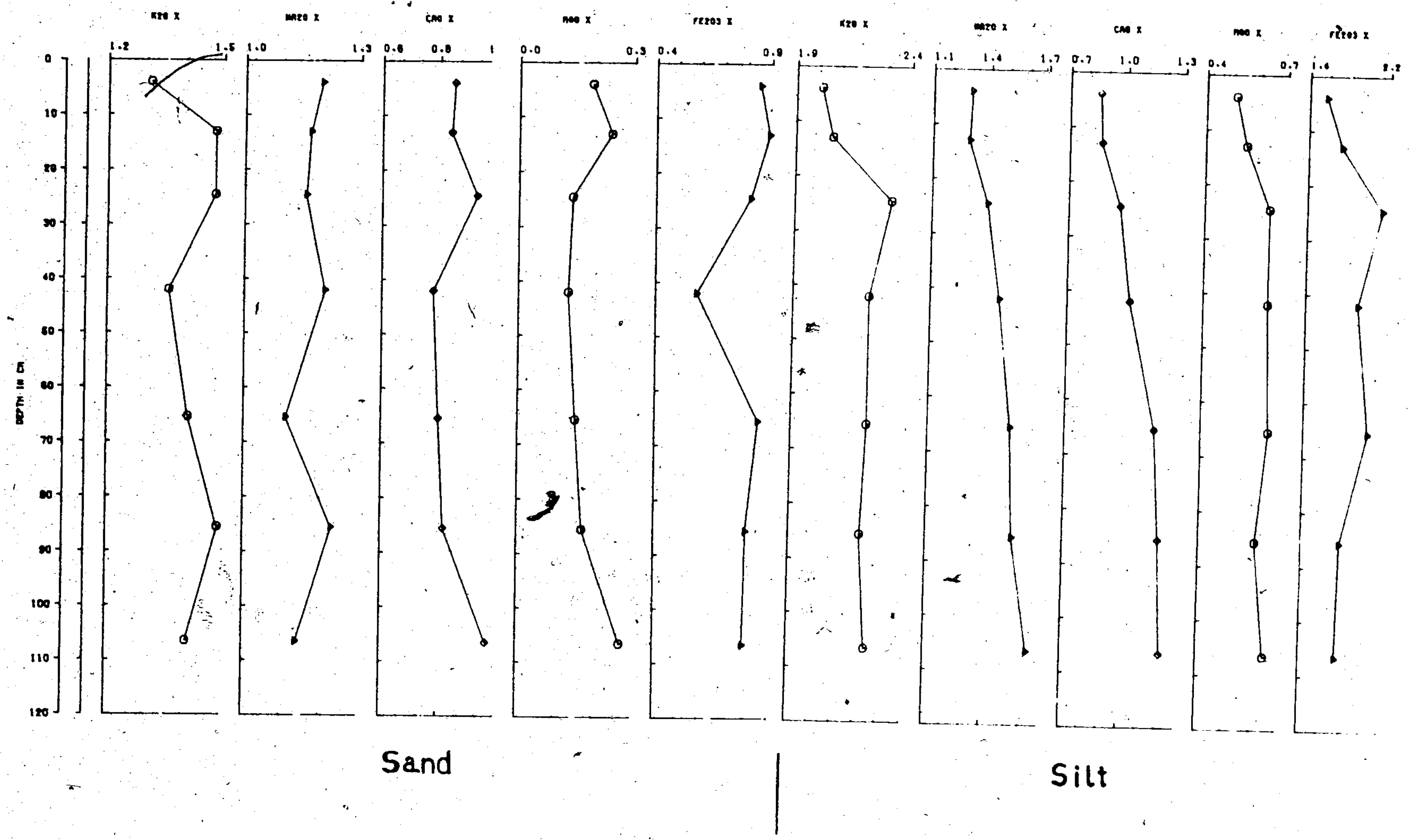


Figure 68. Depth distribution of elemental oxides for a Dark Gray Luvisol (site 3)...................

Table 28. Ratios of potassium feldspars, plagioclases and heavy minerals to quartz in the sand (2-0.02 mm) and silt (0.02-0.002 mm) fractions.

	Mineral ratios x100					
	Sand*			Silt*		
Horizon	KF	Na-CaF	HM	KF	Na-CaF	HM
			Orthic Gray Luvisol (site 1)			
Ae	9.0	13.7	2.6	17.7	31.2	16.9
AB	9.3	14.3	2.0	17.3	28.6	11.3
Bt1	9.1	16.2	2.7	16.5	25.2	16.5
Bt2	9.4	15.4	2.7	17.0	26.9	19.2
BCk	9.9	16.6	3.5	19.6	30.3	23.2
Cca	10.4	16.6	3.7	19.0	30.5	23.8
			Dark Gray Luvisol (site 2)			
Ahe	8.6	14.7	2.5	15.7	21.0	16.6
Ae	8.4	13.5	2.2	14.5	21.4	15.6
Bt	8.7	14.3	3.4	16.7	24.1	18.4
Cca1	10.5	16.2	3.3	16.0	26.8	22.7
Cca2	10.2	15.9	3.0	19.2	28.9	20.5
			Dark Gray Luvisol (site 3)			
Ahe	9.9	15.6	3.1	14.6	20.6	12.1
Ae	10.4	16.2	3.1	15.7	21.5	12.8
AB	10.8	16.6	2.6	17.5	25.1	17.0
Bt1	10.6	18.0	3.2	18.1	24.5	17.6
Bt2	10.5	15.5	3.0	20.4	28.4	19.8
BCk	10.1	15.9	3.2	19.3	28.4	19.8
Cca	10.0	16.7	3.7	19.0	29.2	18.5

* Summation of minerals in the various sand or silt fractions were used to obtain the ratio.

Abbreviations:
KF = K-feldspars, Na-CaF = Na-Calcium feldspars
HM = Heavy minerals

Table 29. Ratios of potassium feldspars, plagioclases and heavy minerals to quartz in the sand (2-0.02 mm) and silt (0.02-0.002 mm) fractions.

	Mineral ratios x100					
	Sand*			Silt*		
Horizon	KF	Na-CaF	HM	KF	Na-CaF	HM
			Orthic Gray Luvisol (site 4)			
Ae	7.4	9.4	2.0	12.2	15.1	12.9
AB	7.5	9.9	1.9	9.0	12.9	15.5
Bt1	7.4	12.0	1.7	13.7	14.8	15.2
Bt2	7.7	13.6	2.0	13.7	16.8	20.4
Ck	7.4	14.5	2.9	13.2	17.9	20.4
			Eluviated Black Chernozemic (site 5)			
Ah	8.1	13.3	2.5	14.9	19.4	16.3
Ae	7.4	11.2	1.7	15.8	19.0	13.1
Bt1	7.3	12.2	2.2	15.7	20.7	15.5
Bt2	6.8	12.3	2.3	16.2	22.7	17.8
BCk	7.9	12.4	2.3	16.7	24.8	20.4
Cca	8.2	14.5	2.7	16.4	27.6	21.2
			Dark Gray Luvisol (site 6)			
Ae	4.9	8.5	1.4	9.5	17.1	11.1
AB	5.1	8.2	1.4	12.8	18.5	13.2
Bt	4.5	7.3	1.4	11.5	16.5	13.0
BCk	4.8	8.5	1.6	12.7	16.7	14.7
Cca	4.7	7.7	2.1	12.7	17.3	14.0

* Summation of minerals in the various sand or silt fractions were used to obtain the ratio.

Abbreviations:
KF = K-feldspars, Na-CaF = Na-Ca feldspars
HM = Heavy minerals

Content of K2O in silt separates invariably increased with profile depth while no regular trends were observed for K2O distribution in sand separates (Figure 68 and Tables 26 and 27). The chemical results agree with the interpretation of mineral ratios where the K-feldspar to quartz ratio for sands show no consistent pattern while the ratio for silts increases with profile depth (Tables 28 and 29) indicative of weathering of K-feldspar in upper solum portions. Since mica also contains potassium, part of the K2O distribution pattern for silts may also refect depotassification of silt-sized mica. The average content of K2O in sand and silt fractions are remarkably similar among soils from sites 1, 2 and 3 and between soils from sites 4 and 5 while content of K2O for the soil from site 6 is almost half the amount found in sand and silt separates from the other 5 soils (Tables 26 and 27). The low K2O content for sands and silts from site 6 soil is matched by a low K-feldspar content (Tables 30 and 31).

Content of Na2O in sand and more commonly in silt separates display a gradual increase with profile depth (Figure 68) which suggests some weathering of Na-bearing minerals in surface horizons. Values for the ratio of plagioclase to quartz (Tables 28 and 29) definitely indicate Na-Ca feldspars in the upper solum have weathered and the profile distribution pattern of Na2O is largely a reflection of abundance of plagioclase feldspar. Trends displayed in the Na2O contents with profile depth are usually matched by a similar trend in the respective feldspar: quartz ratio. Values for both the ratio and Na2O contents for a given profile are either uniform or slightly erratic for sand-sized fractions while silt-sized fractions usually display a progressive increase in both the ratio and content of Na2O with profile depth (excluding the upper horizons of sites 1 and 4 which are now known to be slightly modified or contaminated with eolian material).

Content of CaO generally increases with profile depth for both sand and silt separates (Figure 68 and Tables 26 and 27). The profile distribution pattern of CaO for sand-sized separates depicted in Figure 68 is not typical of the progressive increase with depth displayed by the other five profiles. Like Na2O, the depletion of calcium from upper soil horizons is most pronounced for the more reactive silt-sized fraction and is the result of feldspar weathering as illustrated by the feldspar: quartz ratios. For a given profile, the relative depletion of calcium is greater than the relative depletion of sodium. For the sand-sized fractions, most profiles display progressive and substantial increase in CaO (Tables 26 and 27) with profile depth, however, respective feldspar: quartz ratios are generally erratic and do not match the trend in CaO distribution. These two features suggest that weathering of other mineral constituents, like Ca-bearing pyroxenes and amphiboles may contribute to the profile patterns of CaO. Mineralogical data presented earlier showed that content of amphiboles was lowest in sands and silts from soils at sites 4 and 6. The chemical data for elemental content (Tables 26 and 27) also shows that the lowest content of CaO occurs in sand and silt fractions from soils from sites 4 and 6. Mineral ratios (Tables 28 and 29) definitely indicate heavy minerals have experienced weathering in upper solum horizons. In all likelihood, the CaO distribution patterns for profiles is a reflection of both plagioclase and heavy mineral weathering in surface horizons.

Contents of MgO and Fe2O3 in sand and silt fractions do not show consistant trends indicative of elemental depletion or redistribution despite the fact that heavy mineral: quartz ratios strongly suggest weathering of heavy minerals has occurred in surface and some B horizons. Profile distribution patterns of MgO in silt separates tend to parallel the distribution of vermiculite and chlorite (Tables 30 and 31) while for sand fractions, content of MgO is often lower in the solum compared to C horizons. The rather irregular distribution of total iron in sand and silt fractions within profiles suggests that as heavy minerals weather, the released iron remains as a constituent of a silt or sand-sized pseudomorph or that the released iron forms coatings on sand and silt-sized primary mineral grains. Heavy mineral: quartz ratios clearly indicate that weathering of sand and silt-sized heavy minerals in the solum has occurred in response to pedogenesis.

V. SUMMARY AND CONCLUSIONS

The mineralogical nature of the soils of this study are generally similar. Minerals present in all size fractions are qualitatively similar throughout all sites. The order of abundance of primary minerals in sand and silt fractions is quartz, plagioclase and potassium feldspars. With the exception of substantial amounts of quartz in clay fractions from surface horizons, these primary minerals are present in only minor amounts in the clay fraction. Amphiboles are present in only the sand and silt fractions and concentrate specifically in the coarse and medium silt fractions. Trioctahedral mica is an important component in sand and silt fractions; dioctahedral mica is most abundant in the clay fraction in addition to appearing in the silt fraction. Vermiculite was identified in the silt and clay fractions and is primarily of inherited origin. Smectite, dioctahedral mica and kaolinite were the three main mineral species present in the total clay fraction (<2.0 micrometer) while smectite with subordinate amounts of both hydrous mica and kaolinite were present in the fine clay fraction (<0.2 micrometer). Kaolinite was also present in the fine silt fraction while chlorite was present in the sand, silt and clay fractions. Minerals present in the clay fraction are mainly dioctahedral with minor amounts of trioctahedral minerals interpreted to be biotite and/or trioctahedral vermiculite. Smectite species present in the fine clay fractions are beidellite and montmorillonite as was suggested by the Greene-Kelly test.

The nature of potassium feldspars in the different soils and size fractions are remarkably similar. Potassium feldspars present in the soils of this study are characterized by:

1. the presence of both triclinic and monoclinic polymorphs with average relative proportions of orthoclase equal to 32 ± 6% in the sand fraction and 45 ± 10% in the silt fraction;
2. average relative proportions of 83% $KAlSi_3O_8$, 16% $NaAlSi_3O_8$ and 1% $CaAl_2Si_2O_8$ in the sand fraction and 85% $KAlSi_3O_8$, 13% $NaAlSi_3O_8$ and 2% $CaAl_2Si_2O_8$ in the silt fraction;
3. a perthitic nature. Perthitic intergrowths of albite in potassium feldspars were variable in size and could be classified as macroperthites, microperthites and possibly X-ray perthites.
4. a high degree of exsolution (phase separation) between the two feldspars phases in perthitic potassium feldspars;
5. a high degree of ordering of the microcline polymorph approaching that of maximum microcline as indicated by obliquity values of more than 0.9 Å.

Plagioclase feldspars were also similar between soils and the individual size fractions. Plagioclase species present were dominated by albite or albitic feldspar admixed with lesser amounts of oligoclase and more calcic feldspars. Chemical analysis of plagioclase indicates that on the average it contains 79% NaAlSi308 and 21% CaAl2Si08 in the sand fraction and 82% NaAlSi308 and 18% CaAl2Si208 in the silt fraction. Characterization of the soil plagioclase mixtures by X-ray diffraction methods gave results comparable to those obtained by wet chemical analysis. X-ray diffraction proved to be useful in determining the composition of soil plagioclase mixtures in this study and with few further refinements, it could be an excellent method for qualitative and quantitative analysis of plagioclase species in soils, espcially for size fractions ranging beyond that which is suitable for petrographic examination (e.g., silt-sized material).

Surface morphology of feldspar minerals as seen under the scanning electron microscope indicate alteration has occurred and was evident from solution pitting, alteration to secondary phyllosilicate minerals and other unidentified mineral materials. The extent of alteration as evidenced by surface morphological features appears more pronounced in plagioclase feldspars than in potassium feldspars. Differential weathering of feldspar phases in potassium feldspar crystals was evident in both petrographic and electron microscopic examinations. X-ray and bulk chemical data support the conclusion that the albite phase in perthitic K-feldspars has been differentially weathered.

Although weathering features observed for potassium and plagioclase feldspars were not confined to the biologically active zone of the soil solum (e.g., surface horizons), the overall change in feldspar distribution with depth indicates weathering during pedogenesis had occurred. Weathering trends in K-feldspars are not well defined while those of plagioclase feldspars indicate their weathering in response to pedogenesis in surface horizons. Other minerals experiencing a definite weathering trend causing their depletion in surface horizons are: sand and silt-sized amphiboles, biotite, and chlorite. No definite weathering trends were observed for clay-sized secondary minerals. Soils included in this study are relatively young soils developed in a subhumid cool climate and the changes and/or weathering trends are (as expected) small. The task of obtaining chemical and/or mineralogical weathering trends is also complicated by the fact that the parent material of the soil profiles is glacial till where the inherent characteristics of tills has probably obscured the picture.

Parent material uniformity was confirmed for four of the six sites (sites 2, 3, 5 and 6). Some modification of the surface materials through eolian additions was concluded for site 4, however, at site 1 the surface horizons appear to be depleted of fine particles through wind and/or water action. The established information about parent material uniformity was essential for the interpretation of some ambiguity in mineralogical data.

Quantitative mineralogical data are established for the whole soil and presented in Tables 30 and 31. The data indicate that quartz is the most abundant mineral in the soil followed by heavy minerals (including mica) and plagioclase feldspars. Smectite, kaolinite and potassium feldspars are third in abundance. Clay translocation (especially smectite) is evident from the data, causing a negative enrichment of quartz and feldspars in surface horizons. Kaolinite distribution indicates some translocation to Bt horizons while heavy minerals (minerals with specific gravity greater than 2.72 g/cc) indicate definite weathering in surface horizons despite the negative enrichment which should occur in surface horizons due to clay translocation.

Table 30. Mineralogical constituents of the total soil (sites 1, 2 and 3).

	K-feldspars			Na-Ca feldspar			Quartz			Mica			Amphibol.		Verm		Chlorite		Others		Kaolinite		Smect.
Horizon	Sand	Silt	Clay	Sand	Silt	Clay	Sand	Silt	Clay	Sand	Silt	Clay	Sand	Silt	Silt	Clay	Silt	Clay	Sand	Silt	Silt	Clay	Clay
										% of total soil													
										Orthic Gray Luvisol (site 1)													
Ae	5.60	1.97	0.07	8.48	3.46	0.07	62.01	11.09	0.59	0.21	0.77	00.89	0.23	0.88	0.03	0.20	0.03	0.03	1.27	0.16	0.51	0.89	00.56
AB	4.00	2.32	0.25	6.17	3.83	0.25	43.07	13.40	1.97	0.15	0.64	07.13	0.16	0.69	0.05	2.21	0.02	0.25	0.78	0.13	0.23	6.40	06.15
Bt1	3.39	2.25	0.30	6.01	3.45	0.30	37.21	13.70	1.51	0.21	1.08	09.36	0.22	0.94	0.09	3.02	0.04	0.06	0.60	0.11	0.44	5.44	10.27
Bt2	3.24	2.54	0.31	5.33	4.01	0.31	34.50	14.92	1.54	0.23	1.39	08.90	0.24	1.17	0.12	3.07	0.06	0.06	0.77	0.14	0.57	6.45	10.13
BCk	3.31	2.38	0.35	5.53	3.68	0.35	33.32	12.14	1.39	0.31	1.24	10.76	0.32	1.25	0.12	3.47	0.06	0.35	0.69	0.14	0.79	7.29	10.76
Cca	3.20	2.54	0.35	5.10	4.09	0.35	30.82	13.41	1.06	0.29	1.25	10.24	0.31	1.51	0.18	2.12	0.06	0.35	0.79	0.19	0.97	9.88	10.94
										Dark Gray Luvisol (site 2)													
Ahe	3.03	3.62	0.58	5.22	4.85	0.19	35.41	23.13	3.65	0.17	1.36	04.61	0.17	1.37	0.74	2.30	0.11	0.04	0.87	0.25	0.76	5.57	02.30
Ae	3.96	2.77	0.23	6.36	4.08	0.11	47.24	19.10	1.24	0.21	1.26	03.16	0.22	1.33	0.16	0.34	0.09	0.11	0.85	0.21	0.86	3.96	02.15
Bt	3.20	2.17	0.13	5.23	3.13	0.13	36.58	12.99	1.30	0.25	0.93	08.24	0.26	1.05	0.23	1.94	0.05	0.32	1.05	0.14	0.60	8.42	11.66
Cca1	3.47	2.34	0.63	5.38	3.91	0.32	33.16	14.62	1.59	0.20	1.01	08.24	0.21	1.86	0.16	2.22	0.05	0.31	0.97	0.24	0.72	8.56	09.83
Cca2	[illegible].39	2.92	0.61	5.30	4.39	0.30	33.24	15.18	1.52	0.16	0.97	08.18	0.19	1.77	0.10	2.12	0.05	0.61	1.11	0.21	0.71	7.88	09.09
										Dark Gray Luvisol (site 3)													
Ahe	3.50	4.10	0.38	5.51	5.77	0.25	35.35	28.05	2.29	0.13	1.46	04.06	0.14	1.31	0.34	1.52	0.01	0.03	0.87	0.24	0.50	2.92	01.27
AE	3.74	4.29	0.23	5.79	5.87	0.12	35.87	27.29	1.17	0.13	1.42	03.28	0.14	1.61	0.10	0.70	0.05	0.12	1.00	0.32	0.67	3.63	02.46
AB	3.52	2.38	0.35	5.39	3.42	0.14	32.48	13.61	1.77	0.11	0.92	09.56	0.12	0.97	0.19	3.19	0.04	0.07	0.93	0.17	0.49	7.79	12.39
Bt1	3.42	2.52	0.34	5.81	3.41	0.34	32.32	13.95	1.38	0.15	1.01	09.29	0.16	1.07	0.15	3.78	0.04	0.07	0.78	0.18	0.57	6.88	12.38
Bt2	3.61	2.62	0.33	5.35	3.65	0.33	34.48	12.88	0.99	0.16	1.18	08.58	0.17	1.05	0.14	3.30	0.05	0.07	0.76	0.14	0.69	7.92	11.55
BCk	3.56	2.82	0.29	5.59	4.16	0.29	35.25	14.64	1.43	0.22	1.18	07.72	0.23	1.31	0.14	1.72	0.05	0.29	1.34	0.22	0.68	7.72	09.15
Cca	3.41	2.98	0.28	5.70	4.58	0.28	34.17	15.68	1.13	0.29	1.15	08.20	0.31	1.31	0.16	2.26	0.05	0.06	2.29	0.22	0.67	7.92	08.21

Table 31. Mineralogical constituents of the total soil (sites 3, 4 and 6).

	K-feldspars			Na-Ca feldspar			Quartz			Mica			Amphibol.		Verm		Chlorite		Others		Kaolinite		Smect.
Horizon	Sand	Silt	Clay	Sand	Silt	Clay	Sand	Silt	Clay	Sand	Silt	Clay	Sand	Silt	Silt	Clay	Silt	Clay	Sand	Silt	Silt	Clay	Clay
								% of total soil															
										Orthic Gray Luvisol (site 4)													
Ae	2.34	3.81	0.35	2.99	4.70	0.17	31.71	31.18	2.78	0.04	1.90	04.18	0.05	1.31	0.43	1.22	0.14	0.17	0.67	0.25	1.08	5.22	03.31
AB	1.73	2.33	0.36	2.27	3.31	0.15	22.96	25.77	1.45	0.03	1.93	08.33	0.03	1.18	0.55	4.71	0.14	0.36	0.59	0.20	0.98	8.33	12.31
Bt1	1.65	2.98	0.41	2.69	3.20	0.16	22.28	21.66	0.82	0.02	1.54	12.27	0.04	1.06	0.37	4.91	0.06	0.08	0.59	0.27	0.86	6.95	15.13
Bt2	1.98	2.28	0.42	3.52	2.77	0.17	25.82	16.44	1.27	0.03	1.41	13.57	0.04	1.09	0.54	5.51	0.05	0.09	1.03	0.26	0.77	5.94	15.26
Ck	2.02	2.23	0.39	3.93	3.03	0.16	27.22	16.97	1.57	0.03	1.62	10.58	0.06	1.19	0.34	4.31	0.06	0.39	0.80	0.25	0.90	8.23	13.72
										Eluviated Black Chernozemic (site 5)													
Ah	3.03	3.36	0.38	4.99	4.36	0.38	37.47	22.49	2.44	0.18	1.12	04.51	0.16	0.98	1.30	2.82	0.13	0.19	0.76	0.14	0.73	5.64	02.44
Ae	3.43	2.69	0.18	5.23	3.24	0.18	46.59	17.02	1.43	0.16	1.05	04.48	0.15	0.84	0.16	1.43	0.04	0.18	0.65	0.12	0.73	5.37	04.65
Bt1	3.03	2.05	0.00	5.05	2.71	0.00	41.36	13.10	1.16	0.18	0.93	07.54	0.17	0.82	0.11	2.32	0.04	0.06	0.65	0.12	0.62	7.25	10.73
Bt2	2.81	2.33	0.00	5.07	3.26	0.00	41.16	14.38	0.80	0.20	1.18	07.42	0.18	1.03	0.14	1.86	0.05	0.27	0.75	0.17	0.77	7.16	09.01
BCk	3.17	2.41	0.26	5.01	3.57	0.26	40.28	14.41	1.31	0.16	1.36	07.31	0.13	1.26	0.11	1.57	0.05	0.26	0.85	0.16	0.97	6.26	08.87
Cca	3.06	2.44	0.27	5.41	4.11	0.27	37.27	14.88	1.37	0.21	1.42	07.37	0.17	1.44	0.11	1.37	0.05	0.27	0.89	0.13	1.11	7.64	08.74
										Orthic Gray Luvisol (site 6)													
Ae	2.57	1.86	0.05	4.43	3.35	0.03	52.03	19.61	1.48	0.05	1.05	02.71	0.08	0.72	0.21	0.86	0.09	0.02	0.77	0.11	0.89	4.31	02.83
AB	2.17	2.07	0.00	3.51	2.99	0.00	42.98	16.20	1.06	0.03	0.96	05.57	0.05	0.76	0.22	2.65	0.04	0.27	0.64	0.16	0.70	7.69	09.28
Bt	1.84	1.61	0.00	2.94	2.31	0.00	40.57	13.99	1.34	0.04	0.84	06.72	0.05	0.63	0.20	3.36	0.03	0.34	0.57	0.11	0.67	8.40	13.44
BCk	2.04	1.90	0.00	3.60	2.51	0.00	42.19	14.99	1.45	0.05	1.03	06.09	0.07	0.79	0.20	3.48	0.04	0.29	0.64	0.14	0.81	6.96	10.73
Cca	2.08	2.17	0.00	3.39	2.96	0.00	44.02	17.09	1.67	0.08	1.20	05.74	0.09	0.93	0.05	2.15	0.05	0.24	0.92	0.18	0.89	6.21	07.89

REFERENCES

ACTON, D.F, CROSSON, L.S.(editors) (mimeo) Guidebooks for tours 2,E1,E2 and E3 held in conjunction with the Eleventh Congress of the International Society of Soil Science(Edmonton,1978).

ALEXIADES, C.A., JACKSON, M.L. Quantitative determination of vermiculite in soils. Soil Sci. Soc. Amer. Proc. 29:522-527., 1965.

ALEXIADES, C.A., JACKSON, M.L. Quantitative clay mineralogical analysis of soils and sediments. Clays and Clay Minerals 14:35-53., 1966.

AL-JANABI, A.M., DREW, J.V. Characteristics and genesis of a Sharpsburg-Wymore soil sequence in southern Nebraska. Soil Sci. Soc. Amer. Proc. 31:238-244., 1967.

ALLEN, R.J., HOLE, F.D. Clay accumulation in some Hapludalf as related to calcareous till and incorporated loess on drumlins in Wisconsin. Soil Sci. Soc. Amer. Proc., 32:403-408., 1968.

ARNOLD, R.W., CLINE, M.G. Origin of surficial deposit in soils of eastern Fulton county, New York. Soil Sci. Soc. Amer. Proc. 25:240-243., 1961.

ARSHAD, M.A., PAWLUK, S. Characteristics of some Solonetzic soils in glacial Lake Edmonton basin of Alberta. II. Mineralogy. J. Soil Sci. 17:48-55., 1966.

ATLAS OF ALBERTA. Gov. of Alberta and the University of Alberta .,The university of Alberta Press in association with the Unversity of Toronto Press., 1969.

BARSHAD, I. Chemistry of the soil. In: <u>Chemistry of soil development</u>.Edited by F.E Bear. London, Chapman and Hall Ltd., 1964, pp. 1-71.

BARTH, T.W.F., Feldspars. Wiley-Interscience, New York., 1969. 261 p.

BAYLISS, P., LEVINSON, A., KLOVAN, J.E. Mineralogy of bottom sediments, Hudson Bay, Canada Bull. Can. Petroleum Geology 18:469-473., 1970.

BAYROCK, L.A. Heavy minerals in till of central Alberta. Alta. Soc. Petroleum Geol. Vol. 10, No. 4:171-184., 1962.

BEKE, G.J., ZWARICH, M.A. Chemical and mineralogical characteristics of cutans from B horizons of three Manitoba Soils. Can. J. Soil Sci. 51:221-228., 1971.

BIRD, J.B. The natural landscapes of Canada :a study in regional earth science. Wiley Publishers of Canada limited, Toronto, 1972. pp. 22-48.

BLOOMFIELD, C. The deflocculation of kaolin by tree leaf leachates. Trans. 5th.,Int. Cong. Soil Sci., 2:280-283., 1954.

BORCHARDT, G.A. Montmorillonite and other smectite minerals. In: Minerals in Soil Environments. Edited by J.B. Dixon. Soil Sci. Soc. Amer, Madison, WI., 1977. pp. 293-330.

BORG, I.Y., SMITH, D.K. In: Feldspar minerals: Vol. 1. Crystal structure and physical properties. By J.V Smith . Springer-Verlag, New York., 1974. pp. 210-215.

BOSTOCK, H.S. Physiography subdivisions of Canada In: Geology and economic minerals of Canada, Edited by R.J.W. Douglas. Queen's Printer for Canada, Ottawa, 1970. pp. 10-30.

BOURNE, W.C., WHITESIDE, E.P. A study of the morphology and pedogenesis of a medial Chernozem developed in loess. Soil Sci. Soc. Amer. Proc. 26:484-490., 1962.

BRUNELLE, A., PAWLUK, S., PETERS, T.W. Evaluation of profile development of some Solonetzic Soils of south central Alberta. Can. J. Soil Sci. 56:149-158., 1976.

BROWN, G.(Editor) The X-ray identification and crystal structures. Mineralogical Society, London, 1961, 544pp.

BRYDON, J.E., KODAMA, H., ROSS, G.J. Mineralogy and weathering of the clay in Orthic Podzols and other podzolic Soils in Canada. Trans. nineth Int. Congr. Soil Sci., (Adelaide, 1968) III:41-51., 1968.

BUNDY, L.G., BREMNER J.M. A simple titrimetric method for determination of inorganic carbon in soils. Soil Sci. Soc. Amer. Proc. 36:273-275., 1972.

BYRNE, P.J.S., FARVOLDEN, R.N. The clay mineralogy and chemistry of the Bearpaw formation of southern Alberta. Res. Counc. Alta. Geol. Div. Bull. 4., 1959.

CANN, D.B., WHITESIDE, E.P. A study of the genesis of a Podzol-Gray Brown Podzolic intergrade soil profile in Michigan. Soil Sci. Soc. Amer. Proc. 19:497-501., 1955.

CARRIGY, M.A. Geology of the McMurray Formation part III. Res. Counc. Alta. Memoir 1., 1959.

CARRIGY, M.A. Lithostratigraphy of the uppermost Cretaceous and Paleocene strata of the Alberta plains. Res. Counc. Alta. Bull. 27., 1971.

CARTER, D.L., HEILMAN, M.D., GONZALEZ, C.L. Ethylene glycol monoethyl ether for determining surface area of silicate minerals. Soil Sci. 100:356-360., 1965.

CHAPMAN, S.L., HORN, M.E. Parent material uniformity and origin of silty soils in north west Arkansas based on zirconium:titanium ratios: Soil Sci. Soc. America, Proc. 31: 265-271., 1968.

CHICHESTER, F.W., YOUNGBERG, C.T., HARWARD, M.E. Clay mineralogy of soils formed on Mazama pumice. Soil Sci. Soc. Amer. Proc. 33:115-120., 1969.

CLARK, J.S. and GREEN, A.J. Some characteristics of Gray Soils of low base saturation from northeastern British Columbia. Can. J. Soil Sci. 44:319-328.,,1964.

CLAYTON, J.S., EHRLICH, W.A., CANN, D.B., DAY, J.H., MARSHALL. Soils of Canada, Soil report. The soil Res. Inst., Ottawa, Vol. 1., 1977.

COACH, N.K., KRINSLEY, D.H. Comparison of Strtigraphic and electron microscopic studies in Virginia Pleistocene coastal sediments. J. Geol. 79:426-437., 1971.

COFFMAN, C.B., FANNING, D.S. "Vermiculite" determination on whole soils by cation exchange capacity method. Clays Clay Miner. 22:271-283., 1974.

DE KIMPE, C.R. Weathering of clay minerals in Podzols from the Appalachian highlands Can. J. Soil Sci. 54:395-401., 1974.

DE KIMPE, C.R. Influence of parent material and moisture regime on soil genesis in the Appalachian highlands, Quebec. Can. J. Soil Sci. 56:271-283., 1976.

DEVORE, G.W. The surface chemistry of feldspars as an influence on their decomposition products. Sixth Nat. Conference on Clays and Clay minerals 6:26-41., 1959.

DOUGLAS, L.A. Vermiculites. In: Minerals in Soil Environments. Edited by J.B. Dixon. Soil Sci. Soc. Amer. Inc., Madison, WI., 1977. pp. 259-330.

DREES, L.R., WILDING, L.P. Elemental variability within a sampling unit: Soil Sci. Soc. America Proc. 37:82-87., 1973.

DREIMANIS, A., VAGNERS, U.J. Bimodal distribution of rock and mineral fragments in basal till. In: till: A symposium. Edited by R.P. Goldthwait, Ohio State University Press, Columbs, Ohio. 1971. pp. 237-250.

DUCHAUFOUR, PH. Precis de pedologie. 3rd ed. Masson et Cie, Paris., 1970.

DUDAS, M.J. Clay organo studies of selected Chernozems. Unpublished M.Sc. thesis, Department of Soil Science, U. of A., Edmonton, Alta, 1968.

DUDAS, M.J., PAWLUK, S. Chernozem soils of the Alberta parklands. Geoderma 3:19-36., 1970.

DIJKERMAN, J.C., CLINE, M.G, OLSON, G.W. Properties and genesis of textural subsoil lamellae. Soil Sci. 104:7-16., 1967.

DUDAS, M.J., PAWLUK, S. Naturally occurring organo-clay complexes of Orthic Black Chernozems. Geoderma 3:5-17., 1970.

EDWARDS, A.P., BREMNER, J.M. Dispersion of soil particles by sonic vibration. J. Soil Sci. 18:47-63., 1967.

EHRLICH, W.A., RICE, H.M. Postglacial weathering of Mankato till in Manitoba.Can. J. Soil Sci. 36:527-537., 1956.

ESWARAN, E., BIN, W.C. A study of a deep weathering profile on granite in Peninsular Malaysia.III. Alteration of feldspars. Soil Sci. Soc. Amer. J. 42:154-158.,1978.

EVANS, L.J., ADAMS, W.A. (a) Quantitative pedological Studies on soils derived from Silurian mudstones. IV. Uniformity of the parent material and evaluation of internal Standards. J. Soil Sci., 26:319-326., 1975.

EVANS, L.J. Quantification and pedological processes. In: Quaternary soils. Edited by W.C. Mahaney. Geo. Abstract, Norwich, England, 1978. pp. 361-377.

EXLEY, C.S Some factors bearing on the natural synthesis of clays in granites of southwest England. Clay Minerals Bull. 5:411-426., 1964.

FLINT, R.F. Glacial and Quaternary geology. John Wiley and sons, Inc., New York, 1971. 892 p.

FORMAN, S.A., RICE, H.M. A mineralogical study of some core samples from the Bearpaw formation. Can. J. Soil Sci. 39:178-184, 1959.

FRIEDMAN, G.M., SANDERS, J.E. Principles of sedimentology. John Wiley and sons, New York. 792 pp., 1978.

GIBBS, R.J. Error due to segregation in quantitative clay mineral analysis by X-ray diffraction mounting techniques. Amer. Mineral. 50:741-751., 1965.

GOLDICH, S.S. A study in rock weathering. J. Geol. 46:17-58., 1938.

GOLDSMITH, J.R., LAVES, F. The microcline-sanidine stability relations. Geoch. Cosmo. Acta. 5:1-19., 1954a.

GOLDSMITH, J.R., LAVES, F. Potassium feldspars structurally intermediate between microcline and sanidine. Geoch. Cosmo. Acta. 6:100-118., 1954b.

GOLDTHWAIT, R.P. Introduction to till, today. In: Till: A symposium, Ohio State University Press, Columbus, Oh., 1971. pp. 3-26.

GOLD, C.M. Quantitative methods in the evaluation of the Quaternary geology of the Sand River (73L) map sheet. Unpablished Ph.D thesis, Department of Geology, U. of A., Edmonton, Alta., 1978.

GRAVENOR, C.P., BAYROCK, L.A. Glacial deposits of Alberta In: <u>Soils in Canada</u>. Edited by R.F. Leggett, Rev. Ed., R. Soc. Can. Spec. Publ. 3. University of Toronto Press., 1965. pp. 33-50.

GREENE-KELLY, R. The identification of montmorillonoids in clays. J. Soil Sci. 4:233-237., 1953.

GRIM, R.E. Clay mineralogy. McGraw-Hill, New York. 1968. 596 pp.

HARWARD, M.E., BRINDLEY, G.W. Swelling properties of synthetic smectites in relation to lattice substitutions. Clay and Clay Minerals 13:209-222., 1964.

HARWARD, M.E., CARSTEA, D.D., SAYEGH, A.H. Properties of vermiculites and smectites: expansion and collapse. Clays and Clay Minerals 16:437-447., 1969.

HUANG, P.M., LEE, S.Y. Effect of drainage on weathering transformations of mineral colloids of some Canadian Prairie soils. Proceedings of the Int. Clay Conference Tokyo, 1:541-551., 1969.

HUANG, P.M. Feldspars, olivines, pyroxenes and Amphiboles In:Minerals in soil environments. Edited by J.B. Dixon. Soil Sci. Soc. America. Madison, Wisconsin. U.S.A. 1977 pp. 553-602.

INNES, R.P., The comparative micromorphology of argillic horizons of three Gray Luvisols. Unpublished M.Sc thesis. Department of Soil Science, U. of A, Edmonton, Alta., 1971.

INNES, R.P , Pluth, D.J. Thin section preparation using an epoxy impregnation for petrograpic and microprobe analysis. Soil Sci . Soc. Amer. Proc. 34:483-485., 1970.

JACKSON, M.L. Soil chemical analysis : Advanced course(Revised). Published by the author. Dept. of Soils, University of Wisconsin, Madison.895 pp., 1975.

JOINT COMMITTEE ON POWDER DIFFRACTION STANDARDS. Selected powder diffraction data for minerals. 1st edition, Published by the Joint Committee on Powder Diffraction Standards, 1601 Park Lane, Swarthmore, Pennsylvania 19081, U.S.A., 1974.

KING, R.H., BREWSTER, G.R. Pedogensis in a subalpine environment. In: Quaternary Environments. Edited by W.C. Mahaney., 1974.

KOCAOGLU, S.S., BRUNELLE, A. Reconnaissance soil survey of the Sand River area (73L). The University of Alberta Bull. N. 3315. Alberta Inst. of Pedology Report No. S-74-34, 1975., 83 pp.

KODAMA, H., BRYDON, J.E. Interstratified montmorillonite-mica from subsoils of the Prairie provinces, Western Canada. Clays Clay Mineral. 13:151-173., 1965.

KODAMA, H., BRYDON, J.E. A study of clay minerals in Podzol soils in New Brunswick, eastern Canada. Clay Mineral. 7:295-309., 1968.

KODAMA, H. Clay minerals in Canadian soils: their origin, distribution and alteration. Can. J. Soil Sci. 59:37-58., 1979.

KONONOVA, M.M. Humus of virgin and cultivated soils. In: *Soil components. Vol.1 Organic components*. Edited by J.E. Gieseking. 1975., pp. 475-526.

KINRSLEY, D.H., DONAHUE, J. Environmental interpretation of sand grain surface textures by electron microscopy. Geol. Soc. Amer. Bull. 79:743-748., 1968.

KRINSLEY, D.H., DOORNKAMP, J.C. Atlas of quartz sand surface textures Cambridge University Press. 1973.,91 pp.

LAVES, F. Phase relations of the alkali feldspars. II. The stable and pseudo-stable phase relations in the alkali feldspar system. J. Geol. 60:549-574., 1952.

LAVKULICH, L.M., PAWLUK, S., ODYNSKY, W. Soil profile development in some Peace River soils. Can. J. Soil Sci. 44:188-195., 1964.

LOCKER, J.G. Petrographic and Engineering properties of fine-grain rocks of Central Alberta. Res. Counc. Alta. Bull. 30., 1973.

MacKENZIE, W.S The orthoclase-microcline inversion. Mineral. Mag.

MARFUNIN, A.S. The feldspars. Translated from Russian. 1966. Israel program for scientific translations, Jerusalem., 1962.

MARGOLIS, S.V. Electron microscopy of chemical solution and mechanical abrasion features on quartz sand grains. Sediment. Geol. 2:243-256., 1968.

MARZOLF, J.E. Sand-grain frosting and quartz overgrowth examined by scanning electron microscopy: the Navajo sandstone, Utah. J. Sediment. Petrology 46:906-912., 1976.

McKEAGUE, J.A. Properties and genesis of three members of the uplands catena. Can. J. Soil Sci. 45:63-77., 1965.

McKEAGUE, J.A., BOURBEAU, G.A., CANN, D.B. Properties and genesis of a bisequa soil from Cape Breton Island. Can. J. Soil Sci. 47:101-110., 1967.

McKEAGUE, J.A., St. ARNAUD, R.J. Pedotranslocation: eluvation-illuviation in soils during the Quaternary. Soil Sci. 107:428-434., 1969.

McKEAGUE, J.A., BRYDON, J.E. Mineralogical properties of ten reddish brown soils from the Atlantic provinces in relation to parent materials and pedogenesis. Can. J. Soil Sci. 50:47-55., 1970.

McKEAGUE, J.A., NOWLAND, J.L. BRYDON, J.A., MILES, N.H. Characterization and classification of five soils from eastern Canada having prominently mottled B horizons. Can. J. SOi Sci. 51:483-497., 1971.

McKEAGUE, J.A., MILES, N.M., PETERS, T.W., HOFFMAN, D.W. A comparison of Luvisolic soils from thee regions in Canada. Geoderma 7:49-69., 1972.

McKEAGUE, J.A., MACDOUGALL, J.I., MILES, N.M. Micromorphological, physical, chemical and mineralogical properties of a catena of soils from Prince Edward Island in relation to their classification and genesis. Can. J. Soil Sci. 53:281-295., 1973.

McKEAGUE, J.A. (Editor) Manual on soil sampling and methods of analysis of the Canada Soil Survey Committee. Can. Soc. of Soil Sci., 1978., 212 pp.

MEL'NIKOVA, M.K., KOVENKA, S.V. Effect of physicohical properties of soil on the movement of clay suspensions along the profile. Translated from Pochvovedeniye, 11:45-50., 1974a. Soviet Soil Science. :682-686.

MEL'NIKOVA M.K., KOVENYA, Simulation studies of lessivage. 10th Int. Soil Science Conf. Moscow. VII:600-605., 1974b.

MILLER, F.P., HOLOWAYCHUK, N., WILDING, L.P. Canfield Silt Loam, a Fragiudalf: I. Macromorphological, physical and chemical properties. Soil Sci. Soc. Amer. Proc. 35:319-324., 1971a.

MILLER, F.P., HOLOWAYCHUK, N., WILDING, L.P. Canfield Silt Loam, a Fragiudalf: II. Micromorphology, physical and chemical properties. Soil Sci. Soc. Amer. Proc. 35:324-331., 1971b

MORTON, J.D. Some Geotechnical aspects of late Cretaceous clay shales of the North American plains region: A review of current works and professional papers. Disseration, Imperial College, London, U.K., 100 pp. 1964.

Muir, I.D. The paragesis and optical properties of ternary feldspars. In: Advanced study Inst. of feldspar(Oslo) #2 Feldspar volume. Edited by Olav.J. Christie. 1962. pp 477-492.

MUTWEWINGABO, B. DEKIMPE, C.R., BOURBEAU, G.A., BARIL, R.W., LAJOIE, P. - etude comparative de sols des Laurentides,Quebec. Can. J. Soil Sci. 57:35-45., 1978.

NICHOLSON, H. MARGARET, MOORE, T.R. Pedogenesis in subarctic iron-rich environment: Schefferville, Quebec. Can. J. Soil Sci. 57:35-45., 1978.

OERTEL, A.C. Some observations incompatible with clay illuviation. Trans. 9th., Int., Cong. Soil Sci. :481-487., 1970.

PAWLUK, S. Some Podzol soils of Alberta. Can. J. Soil Sci. 40:1-13., 1960a.

PAWLUK, S. Characteristics of Gray Wooded soils developed from glacial till deposists in the Peace River district and north central regions of Alberta. Transactions of the seventh International Congress of Soil Science, Madison, Wisc., U.S.A. Vol. IV: pp. 314-322., 1960b.

PAWLUK, S. Characteristics of 14 A° minerals in the B horizons of podzolized soils of Alberta. Clays and Clay Minerals XI:74-82., 1963.

PAWLUK, S. Mineralogical composition of some Gray Wooded soils developed from glacial till. Can. J. Soil Sci. 41:228-240., 1961.

PAWLUK, S., LINDSAY, J.D. Characteristics and genesis of Brunisolic soils of northern Alberta. Can. J. Soil Sci. 44:292-303., 1964.

PAWLUK, S. Soil analysis by atomic absorption spectrophotometry. Atomic Absorption Newsletter 6:No. 3., 1967.

PAWLUK, S., PETERS, T.W., CARSON, J. Soils of the Porcupine Hills region of Alberta. Can. J. Soil Sci. 48:77-88., 1968.

PAWLUK, S., BAYROCK, L.A. Some characteristics and physical properties of Alberta tills. Res. Counc. Alta. Bull. 26. 72 pp., 1969.

PAWLUK, S. Characteristics of Fera Eluviated Gleysols developed from acid shales in northwestern Alberta. Can. J. Soil Sci. 51:113-124., 1971.

PAWLUK, S., BREWER, R. Micromorphological and analytical characteristics of some soils from Devon and King Christian Islands, N.W.T. Can. J. Soil Sci. 55:349-361., 1975.

PAWLUK, S., DUDAS, M.J. Reorganization of soil materials in the genesis of an acid Luvisolic Soil of the Peace River Region, Alberta. Can. J. Soil Sci. 58:209-220., 1978.

PETTAPIECE, W.W. The forest-grassland transition In: *Pedology and quaternary research*. Edited by S. Pawluk. University of Alberta, Edmonton. 1969. pp. 103-113.

PETTAPIECE, W.W., ZWARICH, M.A. Micropedological study of a Chernozemic to Gray Wooded sequence of Soils in Manitoba. J. Soil Sci., 21:138-145., 1970.

PETTAPIECE, W.W., PAWLUK, S. Clay mineralogy of soils developed partially from volcanic ash. Soil Sci. Soc. Amer. Proc. 36:616-619., 1972.

PLAS, L. VAN DER. *The identification of detrital feldspars. Developments in sedimentology 6*. Elsevier Publishing Co., New York. p. 305., 1966.

PRICE, T.W., BLEVINS, R.L, BARNHISEL, R.I, BAILY, H.H. Lithological discontinuity in loessial soils of southwestern Kentuky. Soil Sci. Soc. Amer. Proc. 39:94-97.,1975.

REDMOND, C.E., McCLELLAND, J.E. The occurrance and distribution of lime in calcium carbonate solonchacks and associated soils of Eastern North Dakota. Soil Sci. Soc. Amer. Proc. 23:61-65., 1959.

REDMOND, C.E., WHITESIDE, E.P. Some till derived Chernozem soils in eastern north Dakota: II Mineralogy, micromorphology and development. Soil Sci. Soc. Amer. Proc. 31:100-107., 1967.

RICE, H.M., FORMAN, S.A., PATRY, L.M. A study of some profiles from major soil zones in Saskatchewan and Alberta. Can. J. Soil Sci. 39:165-177., 1959.

RITCHIE A., WILDING L.P, HALL, G.F., STAHNKE, G.R. Genetic implications of B horizons in Aqualfs of northeastern Ohio. Soil Sci. Soc. Amer. Proc., 38:351-358., 1974.

ROSTAD, H.P.W., St. ARNAUD, R.J. Nature of carbonate in two Saskatchewan soils. Can. J. Soil Sci. 50:65-70., 1970.

RUTLEDGE, E.M., HOLOWAYCHUK, N., HALL, G.F., WILDING, L.P. Loess in Ohio in relation to several possible source areas: I Physical chemical properties. Soil Sci. Soc. Amer. Proc. 39:1125-1132., 1975.

RUTHERFORD, G.K., SULLIVAN, D.K. Properties and geomorphic relationships of soils developed on a quartzite ridge near Kingston, Ont. Can. J. Soil Sci. 50:419-429., 1970.

SAYEGH, A.H., HARWARD, M.E., KNOX, E.G. Humidity and temperature interaction with respect to K-saturated expanding clay minerals. Amer. Mineralogist 50:490-495., 1965.

SCOTT, J.S. Geology of Canadian tills. In: *Glacial till: An inter-disciplinary study*. Edited by R.F. Legget. Roy. Soc. of Canada spec. Publ. No. 12. Published by the Roy. Soc. of [illegible] in co-operation with the Nat. Res. Coun. of Canada, 197[illegible], pp. 50-66.

SEN, S.K. Potassium content of natural plagioclases and the origin of antiperthites. J. Geol. 67:479-495., 1959.

SMECK, N.E., WILDING, L.P., HOLOWAYCHUK, N. Genesis of argillic horizons in Celina and Morely soils of western Ohio. Soil Sci. Soc. Amer. Proc., 32:550-556., 1968.

SMITH, G.D. Illinois loess. Bull. Univ. Ill. Agric. Exp. Sta. 490:139-184., 1942.

SMITH, J.V. The powder patterns and lattic parameters of plagioclase feldspars I. The soda-rich plagioclases. Min. Mag. 31:47-68, 1956.

SMITH, J.V. *Feldspar minerals: Vol. I. crystal structure and physical properties. Vol. II. Chemical and textural properties*. Springer-Verlag, New York., 1974.

SNEDDON, J.I., LAVKULICH, L.M., FARSTAD, L. The morphology and genesis of some Alpine soils in British Columbia, Canada. II. Physical, Chemical and mineralogical determination and genesis. Soil Sci. Soc. Amer. Proc. 36:104-110., 1972.

SOMASIRI, S., HUANG, P.M. The nature of K-feldspars of a Chernozemic Soil in the Canadian Prairies. Soil Sci. Soc. Amer. Proc. 35:810-815., 1971.

SOMASIRI, S., HUANG, P.M. The nature of K-feldspars of selected soils in the Canadian Prairies. Soil Sci. Soc. Amer. Proc., 37:461-464., 1973.

SOMASIRI, S., HUANG, P.M. The nature of K-micas of selected soils in the Canadian Prairies. 10th. Inter. Soil Sci. Soci. conf., Moscow VII:107-114., 1974.

ST. ARNAUD, R.J., WHITESIDE, E.P. Physical breakdown in relation to soil development J. Soil Sci. 14:267-281., 1963.

ST. ARNAUD, R.J., MORTLAND, M.M. Characteristics of the clay fraction in a Chernozemic to Podzol sequence of soil profiles in Saskatchewan. Can. J. Soil Sci., 44:88-99., 43:336-349., 1963

ST. ARNAUD, R.J., HERBILLON, A.J, Occurrance and genesis of secondary magnesium-bearing calcites in soils. Geoderma. 9:279-298.,1973.

STEWART, R.B., NEALL, V.E., POLLOCK, J.A., SYERS, J.K. Parent material stratigraphy of an Egmont Loam profile, Taranaki, New Zealand, Aust. J. Soil. Res., 15:177-190., 1977.

SUDOM, M.D, ST. ARNAUD, R.J. Use of quartz, zirconium and titanium as indices in pedological studies. Can. J. Soil Sci. 51:385-396.,1971.

THEISEN, A.A., HARWARD, M.E. A paste method for preparation of slides for clay mineral identification by X-ray diffraction. Soil Sci. Soc. Amer. Proc. 26:90-91., 1962.

TWARDY, A.G. A comparative study of glacial till parent material in west central Alberta. Unpublised M.Sc thesis, Department of Soil Science, U. of A. Edmonton, Alta., 1969.

TWARDY, A.G., PAWLUK, S., LINDSAY, J.D. Comparative study of four tills in west-central Alberta. Can. J. Soil. Sci. 54:69-80., 1974.

WRIGHT, T.L., STEWART, D.B. X-ray and optical study of alkali feldspars: I. Determination of composition and structural state from refined unit cell parameters. Amer. Mineral. 53:38-81., 1968.

WRIGHT, T.L. X-ray and optical study of alkali feldspars: II. An X-ray method for determining the composition and structural state from measurement of 2θ values for three reflections. Amer. Mineral. 53:88-104., 1968.

WRIGHT, W.R., FOSS, J.E. Movement of silt-sized particles in sand columns. Soil Sci. Soc. Amer. Proc. 32:446-448., 1968.

YOON, T.N, VANDER PLUYM, H. Buried channels in Edmonton-Lac La Biche -Cold Lake area. Alberta groundwater development., 1974.

APPENDIX A

Site:1 (Series:La Corey) Legal location:NE 14, R 2, Tp. 65, W 4

Well drained Orthic Gray Luvisol developed on strongly calcareous clay loam glacial till; contains some stones up to 20 cm in diameter and rounded pebbles of igneous and metamorphic origin; also contains soft yellowish siltstones and reddish iron concretions. The topography is moderately rolling and the site is located on a north facing slope of a hummocky morainal plain. This land unit covers 25 per cent of the area and is mainly located in the northwestern and westcentral portions of the Sand River Map Sheet (73-1) (Kocaoglu and Brunelle, 1975). The native vegetation consists predominantly of a well established aspen (Populus tremuloides), wild rose (Rosa acicularis), green alder (Alnus crispa), grasses and mosses. A description of the profile follows:

Horizon	Description
LFH	18-0 cm; black (10YR 2/1 m.) semi and highly decomposed organic matter; abundant fine continuous vertical and plentiful medium and coarse roots; abrupt smooth boundary; 13-19 cm thick.
Ae	0-25 cm; light gray (10YR 7/2 m.); weak fine platy; loamy sand, loose to very friable; few very fine and fine vertical and plentiful medium oblique roots; some stones and pebbles; gradual wary boundary; 22-28 cm thick.
AB	25-37 cm; light brownish gray (10YR 6/2 m.); weak medium blocky; sandy clay loam; friable; few medium oblique and few fine vertical continuous roots; few thin clay skins on ped surfaces; some stones and gravels; gradual wavy boundary; 10-14 cm thick.
Bt1	37-62 cm; light gray to gray (10YR 6/1 m.); strong coarse blocky; sandy clay loam, firm, few fine vertical inped and very few medium oblique roots; many moderate thick clay skins in pore and ped surfaces; some stones and pebbles; diffuse wavy boundary; 21-28 cm think.
Bt2	62-87 cm; light brownish gray (10YR 6/2 m.); moderate medium blocky; clay loam; slightly

	firm; few fine inped random roots; common thin clay skins on ped surfaces; some stones and pebbles; clear wavy boundary; 22-30 cm thick.
BCk	87-98 cm; brown to dark brown (10YR 4/3 m.); very weak fine blocky; clay loam; slightly firm; few thin clay skins on ped surfaces; few sand pockets; few common rounded red iron concretions; moderately calcareous fine spots of carbonate accumulation.
Cca	98-125 cm; massive, weakly calcareous common fine spots of carbonate acumulation.

Site:2 (Series:Spedden) Legal location:SE 22, Tp. 62, R 3, W 4

Well drained Dark Gray Luvisol developed on strong to moderately calcareous clay loam glacial till; stones and gravel of igneous origin and soft siltstones are common. The topography of the site is gently undulating and the landform is morainal plain. This land unit covers 23 per cent of the areal (73-L map sheet) extending from the Beaver River Crossing in the east to the ridged morainal plain in the west (Kacoaglu and Brunelle, 1975). The native vegetation consists predominantly of aspen, wild rose, snowberry (*Symphoricarpos albus*), and grasses. A description of the profile follows:

Horizon	Description
Ahe	0-20 cm; gradual change of color from very dark gray brown (10YR 3/2 m.) on top to very dark gray (10YR 3/1 m.) at the bottom; moderate medium granular; loam; very friable; plentiful fine and very fine vertical and abundant coarse and medium oblique and vertical roots; clear wavy boundary; 18 to 22 cm thick.
Ae	20-28 cm; gray (10YR 5/1 m.), light gray (10YR 7/2 d.); moderate medium platy; sandy loam; friable; few fine random and plentiful medium and coarse oblique roots; clear wavy boundary; 6-20 cm thick.
Bt	28-52 cm; dark grayish brown (10 YR4/2 m.), very dark grayish brown (10YR 3/2) staining on ped surfaces; clay loam; compound moderate medium prismatic and strong medium blocky; few medium and coarse inped vertical roots; many moderately thick clay skins on ped and pore surfaces; gradual wavy boundary; 20-28 cm thick.
Cca1	52-87 cm; brown to dark brown (10YR 1/3 m.), massive, carbonates dessiminated throughout the matrix, few medium and coarse vertical roots.
Cca2	87 - 125 cm; massive, carbonate accumulation in coarse viens and medium and fine spots.

Site:3 (Series:Spedden) Legal location:SE 27, Tp. 61, R 2, W 4

Well drained Dark Gray Luvisol developed on moderately calcareous clay loam till; has stones and gravels of granitic origin. The topography of the site is undulating and the landform is morainal plain. The native vegetation consists predominantly of aspen, wild rose, red alder, strawberry, some willow (Salix spp) and grasses. A description of the profile follows:

Horizon	Description
LFH	15-0 cm; black (10YR 2/1 m.) semi and highly decomposed organic matter; abundant fine and very fine continuous random roots; abrupt smooth boundary; 13 to 17 cm thick.
Ahe	0-8 cm; dark gray (10YR 4/1 m.); weak fine platy; loam; very friable; abundant fine and very fine random and plentiful medium oblique roots; clear wavy boundary; 6-10 cm thick.
Ae	8-18 cm; light gray (10YR 7/2 m.); strong medium platy; loam; friable; plentiful fine vertical and abundant coarse and medium oblique roots; gradual wavy boundary; 7-12 cm thick.
AB	18-31 cm; grayish brown (10YR 5/2 m.); strong fine blocky; clay loam; slightly firm; plentiful fine random inped and abundant medium random exped roots; common moderately thick clay skins on ped surfaces; gradual wavy boundary; 10-14 cm thick.
Bt1	31-53 cm; very dark grayish brown (10YR 3/2 m.); compound moderate medium prismatic and strong coarse blocky; clay loam; firm; plentiful fine and medium random inped and exped roots; and many moderately thick and thick clay skins on ped surfaces; diffuse wavy boundary; 20-24 cm thick.
Bt2	53-78 cm; brown to dark brown (10YR 4/3 m.); strong coarse blocky; clay loam; slightly firm; many moderately thick clay skins on pore and ped surfaces; gradual wavy boundary; 24-26 cm thick.

BC	78-93 cm; very dark grayish brown (10YR 3/2 m.); strong medium subangular blocky; sandy clay loam; friable; common thin clay skins on ped surfaces; many pebbles; few (10-15 mm) rounded red iron concretions; 12-17 cm thick.
Ck	93-120 cm; dark yellowish brown (10YR 4/4 m.), massive; weakly calcareous; carbonate accumulated in viens; many pebbles.

Site:4 (Series:Athabaska) Legal location:SW 27, Tp. 64, R 13, W 4

Well drained Orthic Gray Luvisol developed on moderately calcareous clay loam glacial till; dark gray in color and contains few stones. The topography of th site is moderately rolling portion of ridged morainal plain. This land unit covers approximately 5 per cent of the area in a narrow strip in the western part of the map sheet(73L) extending from northwest to southeast (Kocaoglu and Brunelle, 1975). The native vegetation consisted predominantly of aspen, wild rose and some understory grasses. A description of the profile follows:

Horizon	Description
LFH	8-0 cm; very dark grayish brown (10YR 3/1 m.); semi-and highly decomposed organic matter; abundant fine continuous random roots; abrupt smooth boundary; 6-10 cm thick.
Ae	0-15 cm; light gray (10YR 7/2 m.); strong coarse platy; loam; friable; abundant fine and medium continuous vertical and plentiful coarse oblique roots; diffuse wavy boundary; 1-3 cm thick.
AB	15-30 cm; dark grayish brown (10YR 4/2 m.); strong fine subangular blocky; clay loam; slightly firm; abundant fine and medium vertical and plentiful coarse oblique roots; many thick clay skins on ped and pore surfaces; gradual irregular boundary; 13-17 cm thick.
Bt1	30-71 cm; very dark brown (10YR 3/2 m.); compound moderate medium prismatic and strong fine blocky; clay loam; firm; abundant medium and fine exped and inped roots and few coarse oblique exped roots; continuous very thick clay skins on ped surfaces; diffuse wavy boundary; 39-43 cm thick.
Bt2	71-98 cm; very dark grayish brown (10YR 3/2 m.); compound weak medium prismatic and moderate medium blocky; clay loam; firm; abundant medium and fine vertical exped and inped roots; many thick clay skins on ped surfaces; common (5-20 mm) rounded red iron concretions; diffuse wavy boundary; 25-30 cm thick.

Ck 98-124 cm; dark grayish brown (10YR 4/2 m.); compact; massive; few-common (5-20 mm) rounded red iron concretions; few soft yellow silt stones; weakly calcareous.

Site:5 (Series:Fergy). Legal location:SW Sec 13, Tp. 61, R. 6, W 4

Well drained Eluviated Black Chernozemic developed on a moderately high calcareous laom to sandy loam glacial till; contains some stones of granite and highly weathered mica schist and few yellowish siltstones. The topography of the site is gently undulating and the landform is morainal plain. This land unit covers approximately 37 per cent of the map area(73L) located in the north of Beaver river and east of the Sand river as well as an area south of Wolf Lake (Kacaoglu and Brunelle, 1975). The native vegetation consisted of brome grass, chokecherry (Prunus virginia), wild rose and small shrubs and aspen trees. The description of the profile follows:

Horizon	Description
Ah	0-23 cm; black (10YR 2/1 m.); strong fine gr abundant very fine and fine random roots; abrupt wavy boundary; 21-25 cm thick.
Ae	23-32 cm; light gray (10YR 7/2 m.); weak fine platy; sandy loam; friable; abundant very fine and fine random roots; clear wavy boundary; 6-10 cm thick.
Bt1	31-59 cm; brown to dark brown (10YR 4/3 m.); weak medium and fine blocky; sandy clay loam; slightly firm; plentiful fine random and vertical inped roots; few thin clay skins on ped surfaces; gradual diffuse boundary; 26-31 cm thick.
Bt2	59-83 cm; brown (10YR 5/3 m.); dark yellowish brown (10YR 3/4 m.) on ped surfaces; compound moderate coarse prismatic and moderate medium sub angular blocky; sandy clay loam; firm; common fine random inped roots; few very thin clay skins on ped surfaces; clear wavy boundary; 12-16 cm thick.
BCk	83-100 cm; yellowish brown (10YR 5/4 m.); weak fine subangular blocky sandy clay loam; friable; few fine random roots; few stones 5-10 cm in diameter; few (5-15 mm) rounded red iron concretions; weakly calcareous; clear wavy boundary; 15-20 cm thick.

Cca 100-125 cm; dark grayish brown (10YR 4/2 m.); strongly to moderately calcareous; splotched with white carbonate accumulations; few stones of granite and mica schist; few (5-20 mm) rounded red iron concretions.

Site:6 (Series:Cooking Lake Loam) Legal location:NW Sec 14, Tp. 52, R 23, W 4

Welldrained Orthic Gray Luvisol developed on moderately calcareous sandy clay loam glacial till; contains stones of granitic origin. The topography of the site is moderarely rolling and is located on hummocky moraine. The native vegetation consists of aspen, wild rose hazel and grasses. The description of the profile follows:

Horizon	Description
LF	2-0 cm; semidecomposed organic matter; abrupt smooth boundary; 6-9 cm thick.
Ae	0-15 cm; light gray (10YR 7/1 d.) sandy loam; strong coarse and medium platy; soft; few coarse horizontal exped and plentiful fine vertical inped roots; clear smooth boundary 10-16 cm thick.
AB	15-27 cm; brown (10YR 4/3 m.) sandy clay loam; strong; fine and medium subangular blocky; firm; few medium random exped and few vertical inped and exped roots; few very fine clay skins; clear smooth boundary; 10-15 cm thick.
Bt1	27-40 cm; dark brown (10YR 3/3 m.) sandy clay loam; weakly prismatic and strong fine to medium subangular blocky; very firm; few coarse lateral and few fine vertical inped and exped roots; continuous moderately thick clay skins; clear smooth boundary; 10-16 cm thick.
Bt	40-65 cm; dark grayish brown (10YR 4/2 m.); sandy clay loam; strong fine to medium subangular blocky; firm; few medium random and few fine vertical inped and exped roots; few fine clay skins; gradual smooth boundary; 23-28 cm thick.
BC	65-75 cm; greyish brown (10YR 5/2 to 5/3 m.) sandy clay loam; fragmental and strong fine subangular blocky; firm; few fine vertical exped and inped roots; few fine clay skins; clear wavy boundary; 7-10 cm thick.

Cca	75+ cm; dark greyish brown (10YR 4/2 cm) sandy clay loam; friable; moderately calcareous till.

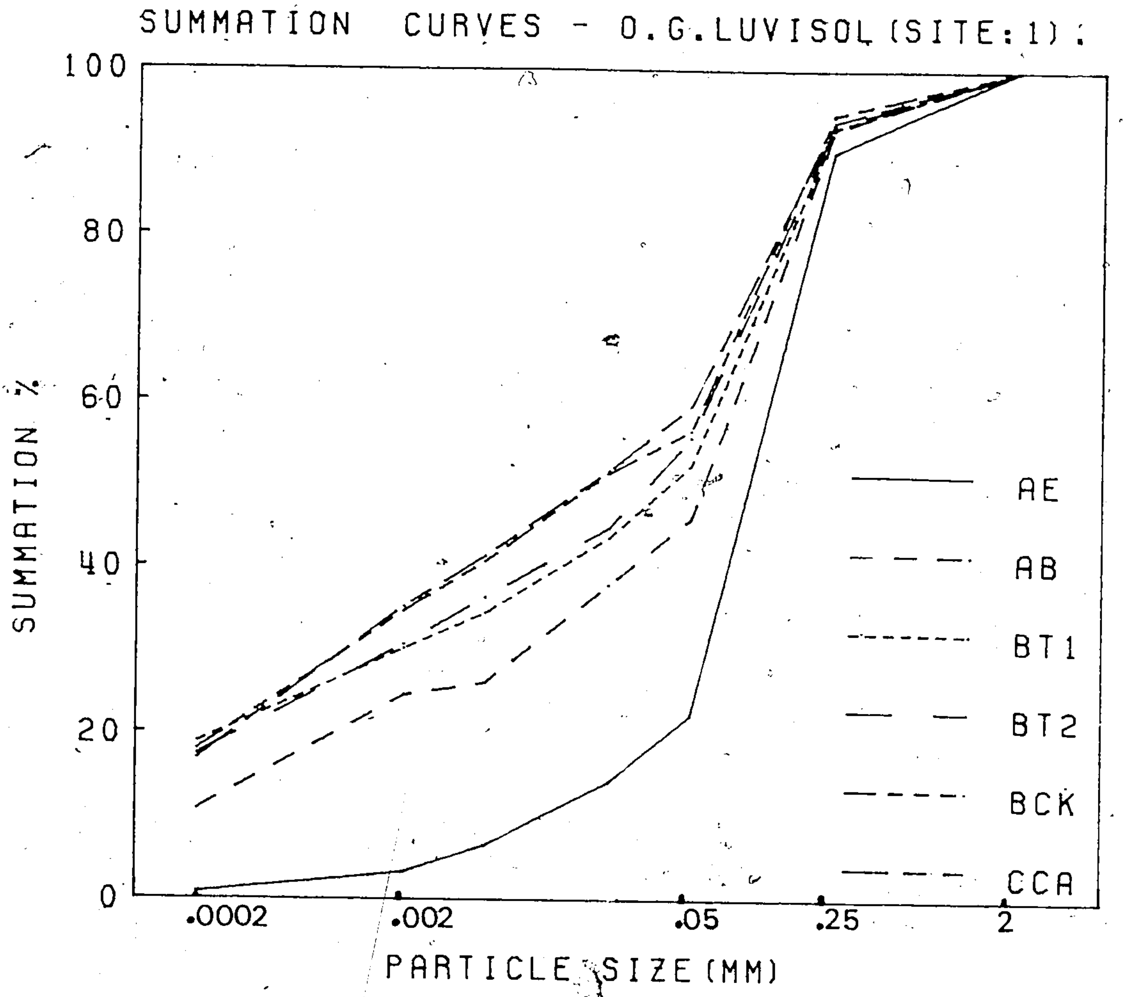


Figure 1. Particle size cumulative-frequency curves representing horizons at site 1...............

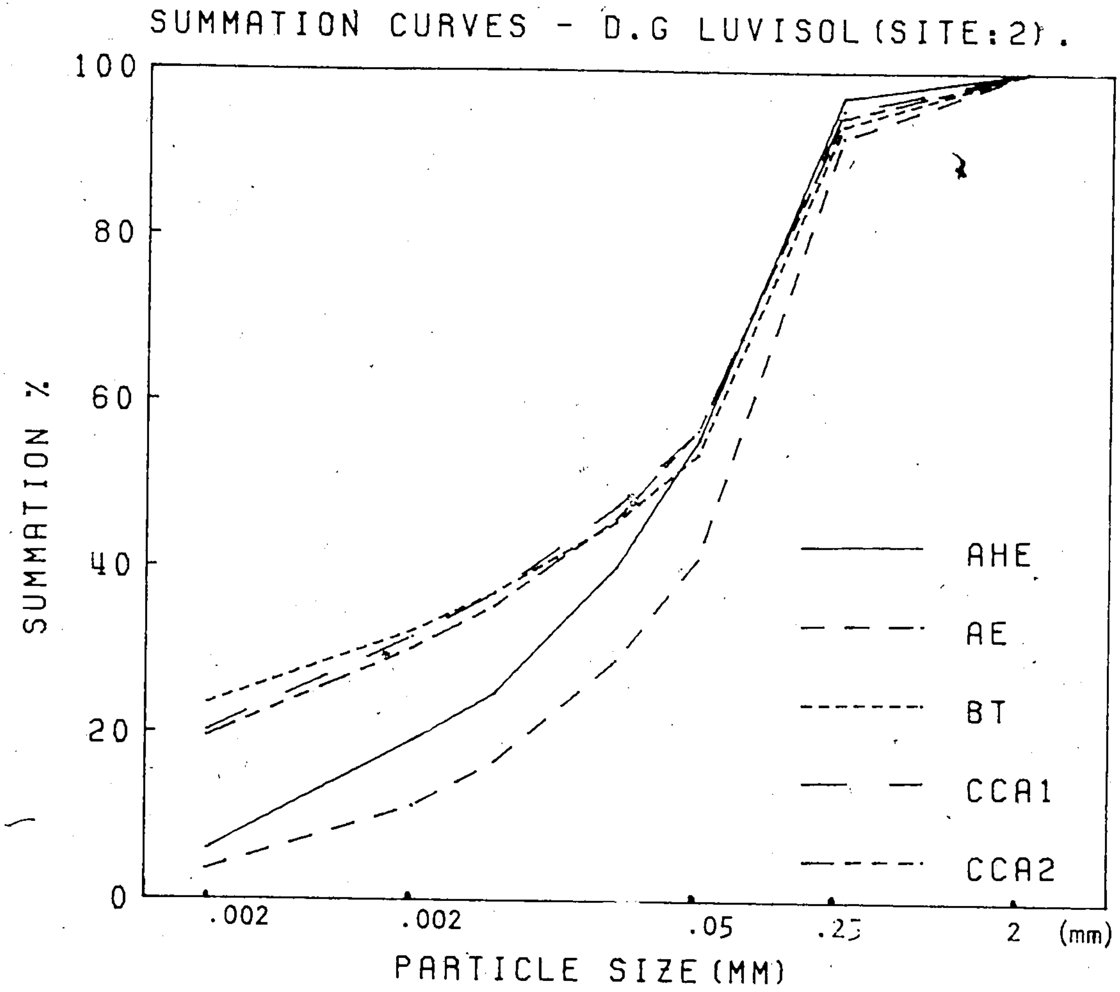


Figure 2. Particle size cumulative-frequency curves representing horizons at site 2.............. .

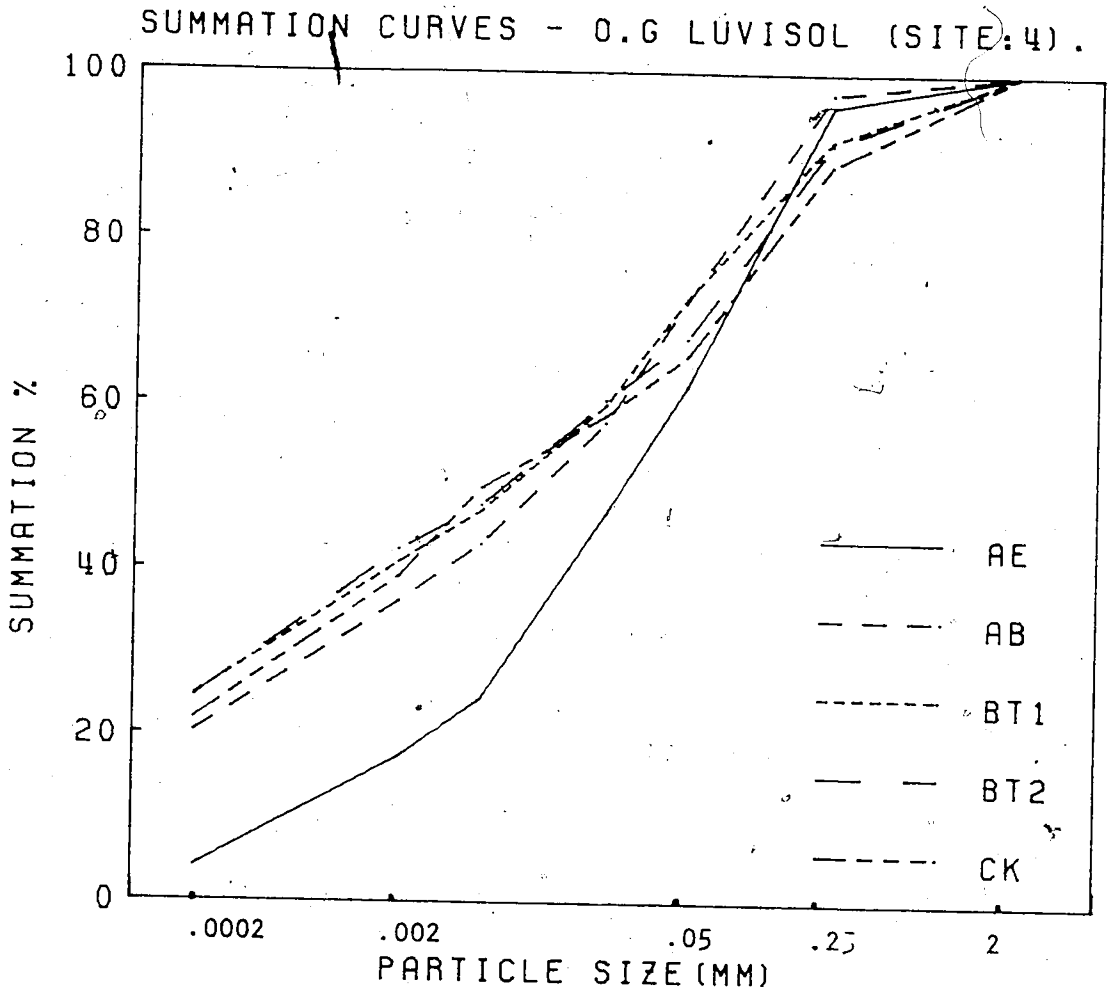


Figure 3. Particle size cumulative-frequency curves representing horizons at site 4................

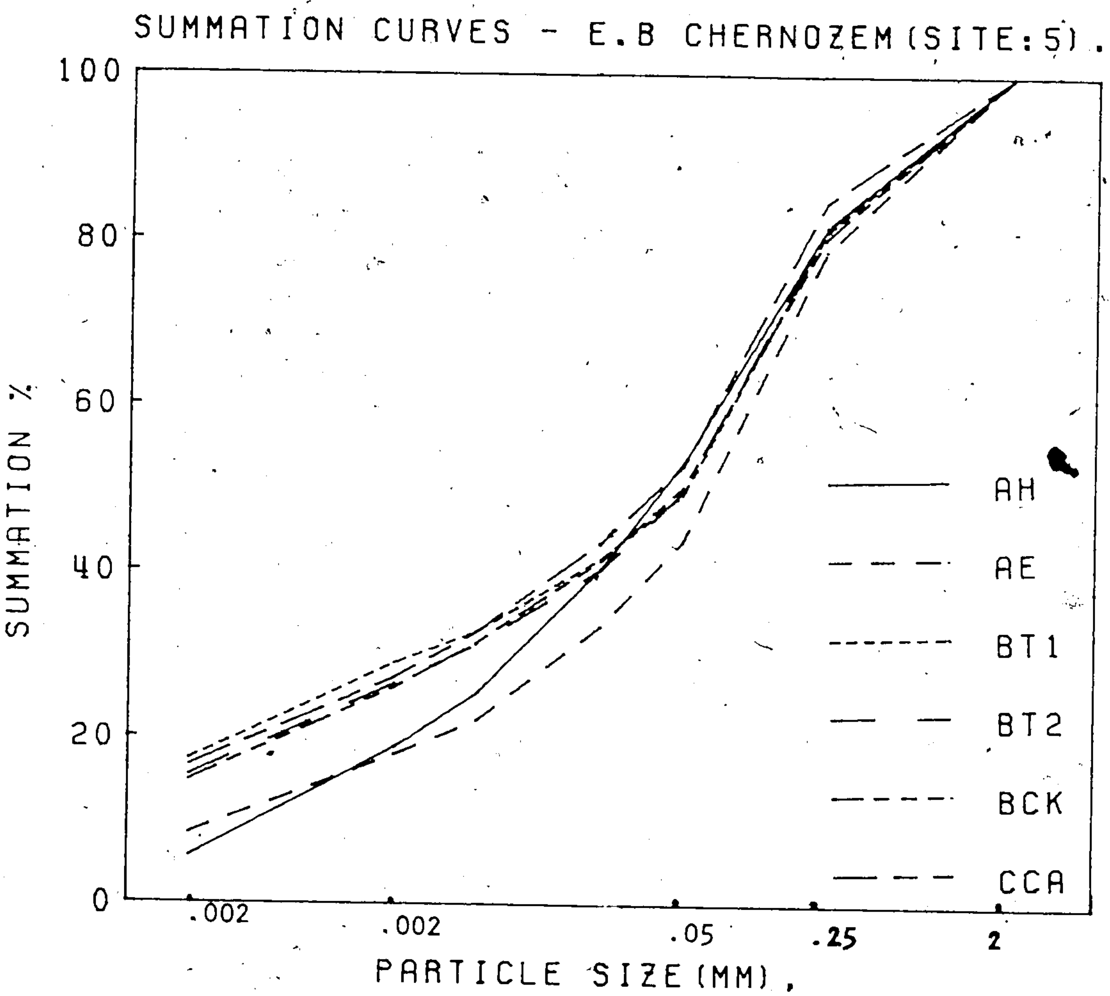


Figure 4. Particle size cumulative-frequency curves representing horizons at site 5..............

Table 1. Chemical and physical soil properties.

Site		1					
Classification		Orthic Gray Luvisol					
Horizon		Ae	AB	Bt1	Bt2	BCk	Cca
Horizon depth		0-25	-37	-62	-87	-98	-125
pH in distilled H20		6.60	5.80	5.60	5.90	6.30	7.90
pH in 0.01 N CaCl2		5.70	5.10	4.90	5.00	5.50	7.50
Organic matter %		0.47	0.65	0.78	0.95	0.96	0.96
CaCO3 equivalent		0.00	0.00	0.00	0.00	0.10	5.80
Exchangeable cations	Al&H*	0.72	1.94	3.90	4.10	ND	ND
	Na	0.11	0.10	0.02	0.03	0.04	0.04
	K	0.16	0.45	0.59	0.59	0.59	0.41
	Ca	1.44	5.78	3.16	8.25	ND	ND
	Mg	0.27	2.03	7.45	3.53	ND	ND
C.E.C. me/100g		2.70	10.30	15.20	16.50	17.70	15.70
Particle size analysis**							
C.&m sand %		9.80	7.00	6.9	6.50	5.30	6.1
F. sand %		68.00	47.10	40.8	37.90	38.20	34.4
C. silt %		8.10	8.00	8.7	10.80	5.10	7.8
M. silt %		7.60	11.80	9.1	8.40	10.70	10.3
F. silt %		3.2	1.5	4.3	5.7	6.0	6.1
C. clay %		2.5	13.8	11.3	13.3	16.7	18.4
F. clay %		0.8	10.8	18.9	17.4	18.0	16.9
F.sand:T.sand ratio		0.87	0.87	0.86	0.85	0.88	0.85
Silt:Clay ratio		5.72	0.87	0.73	0.81	0.63	0.69
F.clay:C.clay ratio		0.32	0.78	1.67	1.31	1.08	0.92

* Calculated by difference

** Mean of two replicates while others are mean of three replicates

Table 2. Chemical and physical soil properties.

Site		2				
Classification		Dark Gray Luvisol				
Horizon		Ahe	Ae	Bt	Cca1	Cca2
Horizon depth		0-20	-28	-52	-87	-125
pH in distilled H20		6.30	6.30	7.20	8.40	8.40
pH in 0.01 N CaCl2		5.60	5.30	6.20	8.00	8.00
Organic matter %		9.28	1.19	1.29	0.71	0.81
CaCO3 equivalent		0.00	0.00	0.00	0.00	0.10
Exchangeable cations	Al&H*	5.42	2.03	2.22	ND	ND
	Na	0.11	0.10	0.11	0.21	0.19
	K	0.26	0.10	0.67	0.19	0.17
	Ca	9.29	1.89	4.23	ND	ND
	Mg	6.22	2.48	12.87	ND	ND
C.E.C. me/100g		21.30	6.60	20.10	15.70	15.00
Particle size analysis**						
C.&m sand %		3.10	8.00	6.5	4.70	5.50
F.sand %		41.50	50.90	39.8	38.70	37.90
C. silt %		15.20	12.20	8.1	9.20	10.60
M. silt %		15.30	12.30	8.6	10.50	10.60
F. silt %		5.7	5.3	4.6	5.2	5.1
C. clay %		13.1	7.7	8.8	11.4	10.7
F. clay %		6.1	3.6	23.6	20.3	19.6
F.sand:T.sand ratio		0.93	0.86	0.86	0.89	0.87
Silt:Clay ratio		1.89	2.64	0.66	0.79	0.87
F.clay:C.clay ratio		0.47	0.47	2.68	1.78	1.83

* Calculated by difference

** Mean of two replicates while others are mean of three replicates

Table 3. Chemical and physical soil properties.

Site		3						
Classification		Dark Gray Luvisol						
Horizon		Ahe	Ae	AB	Bt1	Bt2	BCk	Cca
Horizon depth		0-8	-18	-31	-53	-78	-93	-120
pH in distilled H20		7.10	7.10	6.90	6.80	6.80	7.20	8.10
pH in 0.01 N CaCl2		6.50	6.40	6.30	6.30	6.00	6.70	7.50
Organic matter %		6.23	1.20	1.53	1.29	1.15	0.88	0.74
CaCO3 equivalent		0.00	0.00	0.00	0.00	0.00	0.70	6.80
Exchangeable cations	Al&H*	1.92	1.49	3.30	1.84	1.81	ND	ND
	Na	0.01	0.01	0.01	0.02	0.03	0.05	0.05
	K	0.64	0.36	0.60	0.37	0.32	0.31	0.34
	CA	11.15	4.13	12.13	11.73	9.42	ND	ND
	Mg	1.68	0.91	5.06	5.54	5.32	ND	ND
C.E.C. me/100g		15.40	6.90	21.10	19.50	16.90	14.80	13.50
Particle size analysis**								
C.&m sand %		3.20	6.40	5.4	5.20	5.60	5.1	4.8
F. sand %		42.30	40.30	37.0	37.50	39.00	41.1	40.1
C. silt %		17.60	17.60	6.6	7.80	8.40	7.7	10.1
M. silt %		18.50	19.00	11.8	11.30	9.30	12.7	11.5
F. silt %		5.7	5.0	3.8	3.8	4.7	4.8	5.2
C. clay %		10.2	8.1	13.4	14.5	12.8	10.8	11.8
F. clay %		2.5	3.6	22.0	19.9	20.2	17.8	16.5
F.sand:T.sand ratio		0.93	0.86	0.87	0.88	0.87	0.89	0.89
Silt:Clay ratio		3.29	3.56	1.02	0.67	0.68	0.88	0.95
F.clay:C.clay ratio		0.25	0.44	1.64	1.37	1.58	1.65	1.40

* Calculated by difference

** Mean of two replicates while others are mean of three replicates

Table 4. Chemical and physical soil properties.

Site		4				
Classification		Orthic Gray Luvisol				
Horizon		Ae	AB	Bt1	Bt2	Ck
Horizon depth		0-15	-30	-71	-98	-124
pH in distilled H2O		6.10	5.20	4.80	6.30	8.00
pH in 0.01 N CaCl2		5.20	4.40	4.30	5.90	7.60
Organic matter %		1.91	1.96	1.75	1.81	1.29
CaCO3 equivalent		0.00	0.00	0.00	0.00	3.30
Exchangeable cations	Al&H*	0.47	3.25	3.22	1.64	ND
	Na	0.03	0.08	0.13	0.13	0.13
	K	0.27	0.29	0.30	0.31	0.30
	CA	4.77	10.92	11.67	13.69	ND
	Mg	1.56	6.26	6.98	7.73	ND
C.E.C. me/100g		7.10	20.80	22.30	23.50	21.30
Particle size analysis**						
C.&m sand %		3.80	2.40	8.0	8.40	11.00
F. sand %		34.00	25.00	19.1	23.60	23.20
C. silt %		15.40	14.90	13.0	7.50	7.10
M. silt %		22.30	14.60	12.8	12.70	13.80
F. silt %		7.1	6.9	6.2	5.4	5.7
C. clay %		13.2	15.8	16.1	17.8	17.2
F. clay %		4.2	20.4	24.8	24.6	22.0
F.sand:T.sand ratio		0.90	0.91	0.71	0.74	0.68
Silt:Clay ratio		2.58	0.65	0.68	0.57	0.77
F.clay:C.clay ratio		0.32	1.29	1.54	1.39	1.28

* Calculated by difference

** Mean of two replicates while others are mean of three replicates

Table 5. Chemical and physical soil properties.

Horizon	AH	Ae	Bt1	Bt2	BCk	Cca
Site	5					
Classification	Eluviated Black Chernozemic					
Horizon depth	0-23	-31	-59	-83	-100	-125
pH in distilled H20	6.90	7.10	6.50	6.80	8.10	8.30
pH in 0.01 N CaCl2	6.40	6.50	5.90	6.10	7.60	7.60
Organic matter %	16.86	1.50	1.32	1.19	0.81	0.77
CaCO3 equivalent	0.00	0.00	0.00	0.00	6.10	8.60
Exchangeable cations Al&H*	0.89	0.62	2.26	1.06	ND	ND
Na	0.02	0.03	0.05	0.11	0.12	0.12
K	1.09	0.18	0.26	0.28	0.29	0.29
CA	20.06	5.21	7.23	6.95	ND	ND
Mg	5.54	2.96	5.48	5.40	ND	ND
C.E.C. me/100g	27.60	9.00	15.20	13.80	11.90	13.30
Particle size analysis**						
C.&m sand %	18.20	21.10	18.7	18.10	19.40	15.1
F. sand %	28.40	35.10	31.8	32.10	30.20	31.9
C. silt %	12.90	10.30	7.8	9.40	8.90	9.3
M. silt %	15.20	11.40	8.9	9.10	10.00	10.9
F. silt %	6.5	4.2	3.8	4.8	5.4	5.5
C. clay %	13.2	9.5	11.7	11.1	11.3	10.7
F. clay %	5.6	8.4	17.3	15.4	14.8	16.6
F.sand:T.sand ratio	0.61	0.63	0.63	0.64	0.61	0.68
Silt:Clay ratio	1.84	1.45	0.71	0.88	0.93	0.94
F.clay:C.clay ratio	0.42	0.88	1.49	1.39	1.31	1.55

* Calculated by difference

** Mean of two replicates while others are mean of three replicates.

Table 6. Chemical and physical soil properties.

Site :	6				
Classification :	Orthic Gray Luvisol				
Horizon :	Ae	AB	Bt1	BCk	Cca
Horizon depth :	0-15	-27	-65	-75	-130
pH in distilled H20:	5.70	5.50	6.50	7.80	8.30
pH in 0.01 N CaCl2 :	5.00	5.00	6.00	7.50	7.80
Organic matter % :	0.95	0.95	0.96	0.88	0.62
CaCO3 equivalent :	0.00	0.00	0.00	1.70	9.70
Exchangeable cations Al&H*:	2.17	2.73	1.43	ND	ND
Na:	0.02	0.06	0.09	0.09	0.12
K:	0.24	0.35	0.21	0.17	0.17
CA:	3.25	9.41	13.51	ND	ND
Mg:	1.32	4.35	6.36	ND	ND
C.E.C. me/100g :	7.00	16.90	21.60	19.20	12.10
Particle size analysis**					
C.&m sand % :	23.00	18.60	16.8	17.20	18.90
F. sand % :	36.80	30.80	29.2	31.40	31.70
C. silt % :	12.00	10.70	8.3	8.90	10.60
M. silt % :	11.60	9.70	8.7	9.50	10.40
F. silt % :	4.3	3.7	3.4	4.0	4.5
C. clay % :	8.2	10.7	12.4	12.0	8.6
F. clay % :	4.1	15.8	21.2	17.1	15.3
F.sand:T.sand ratio:	0.62	0.62	0.64	0.65	0.53
Silt:Clay ratio :	2.27	0.88	0.61	0.77	0.84
F.clay:C.clay ratio:	0.50	1.48	1.71	1.43	1.78

* - Calculated by difference

** Mean of two replicates while others are mean of thre replicates

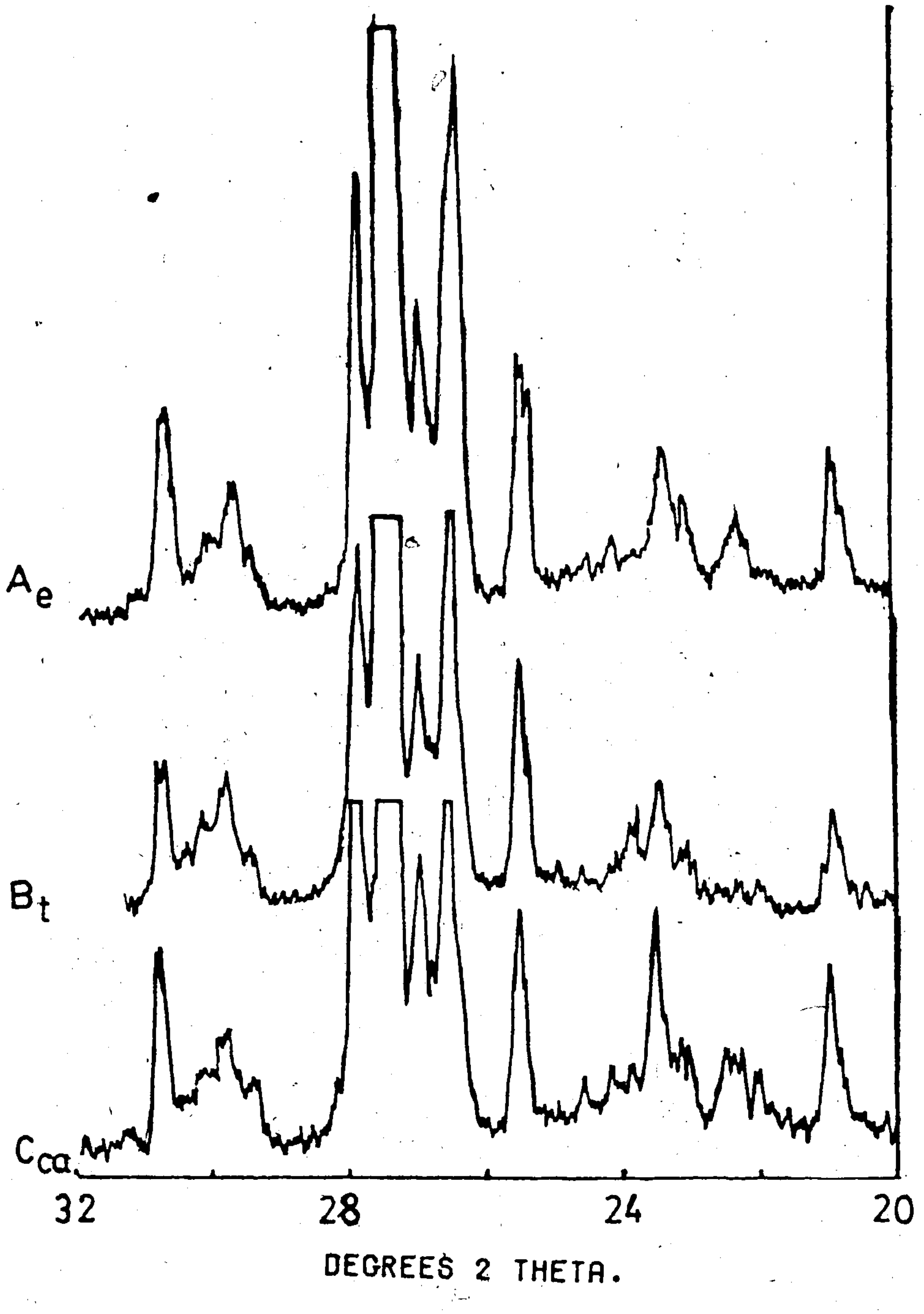


Figure 5. X-ray diffraction patterns for the 2.53 - 2.59 g/cc specific gravity separate from the fine sand fraction of the major genetic horizons from an Orthic Gray Luvisol (site 1) (for peak indices refer to the text).......

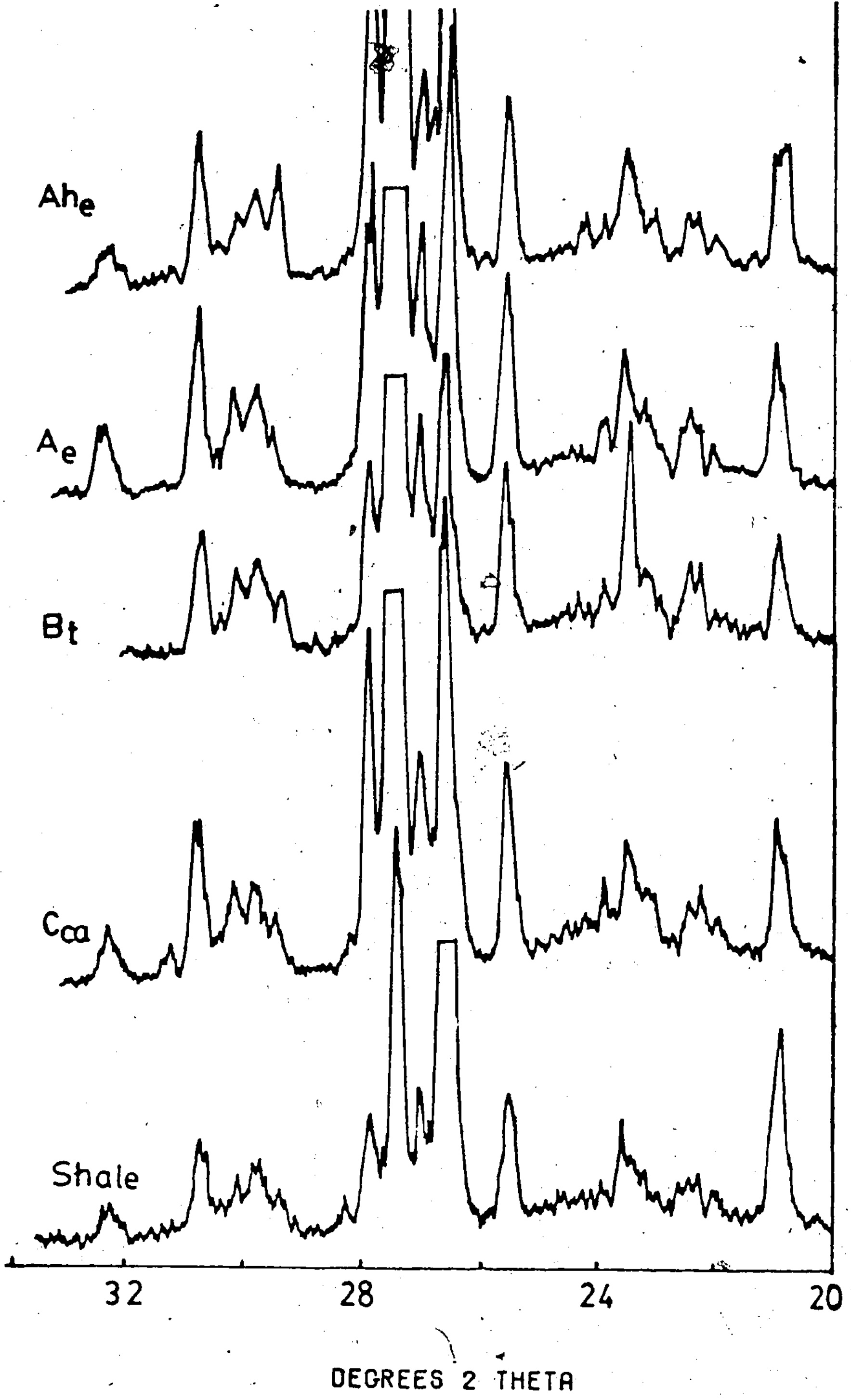


Figure 6. X-ray diffraction patterns for the 2.53 - 2.59 g/cc specific gravity separate from the fine sand fraction of the major genetic horizons from a Dark Gray Luvisol (site 2) (for peak indices refer to the text)..........

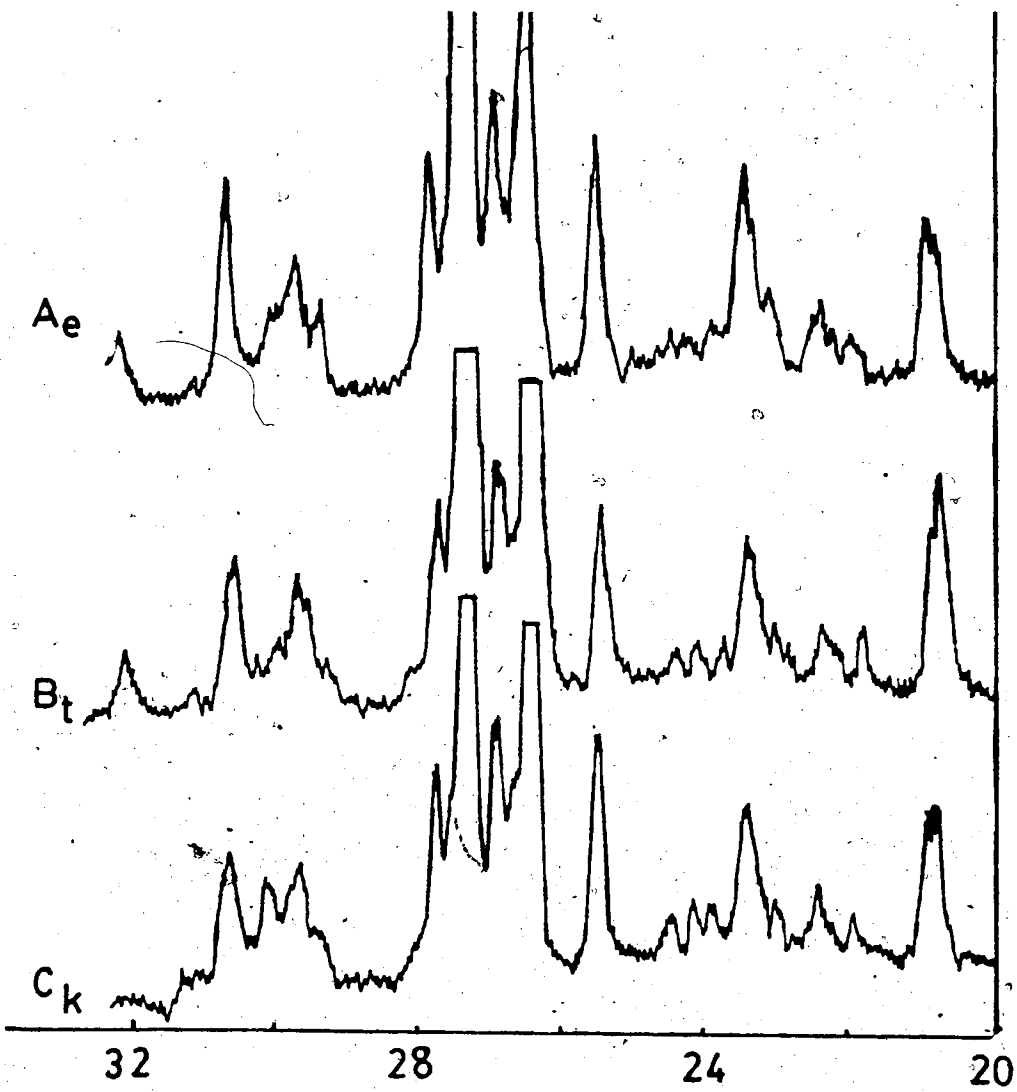


Figure. 7. X-ray diffraction patterns for the 2.53 - 2.59 g/cc specific gravity separate from the fine sand fraction of the major genetic horizons from an Orthic Gray Luvisol (site 4) (for peak indices refer to the text).......

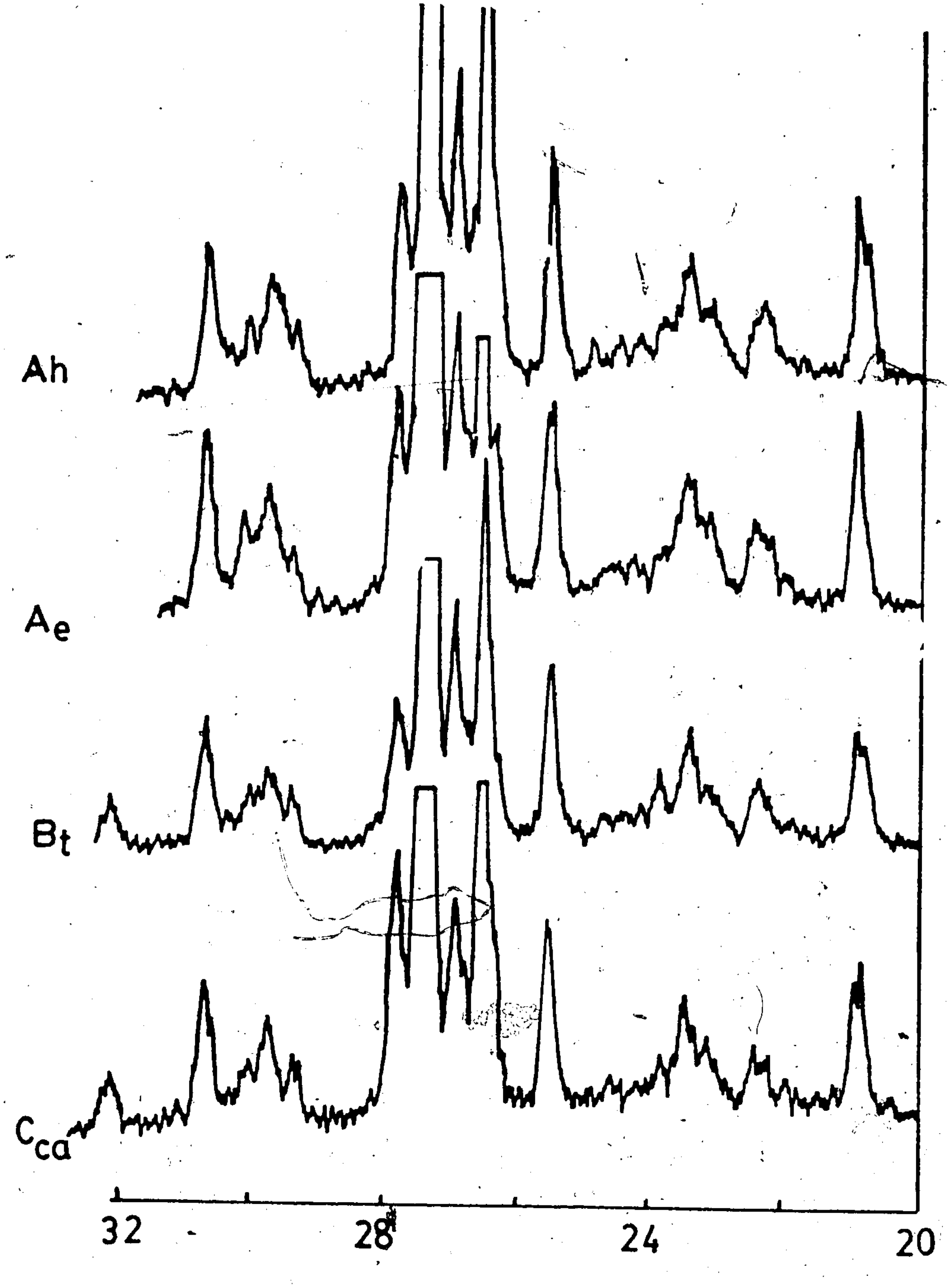


Figure 8. X-ray diffraction patterns for the 2.53 – 2.59 g/cc specific gravity separate from the fine sand fraction of the major genetic horizons from the Eluviated Black Chernozemic (site 5) (for peak indices refer to the text)...

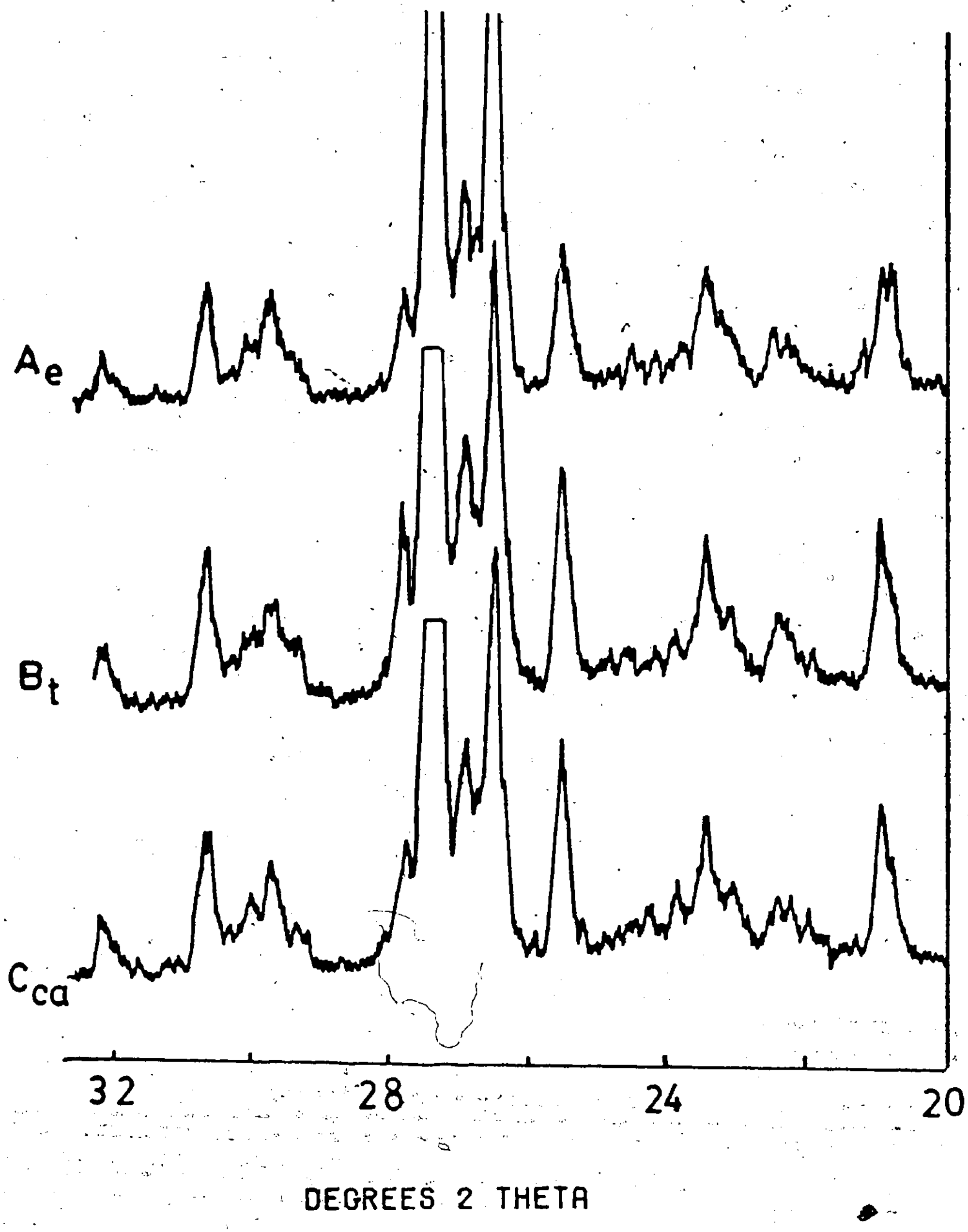


Figure 9. X-ray diffraction patterns for the 2.53 – 2.59 g/cc specific gravity separate from the fine sand fraction of the major genetic horizons from an Orthic Gray Luvisol (site 6) (for peak indices refer to the text).......

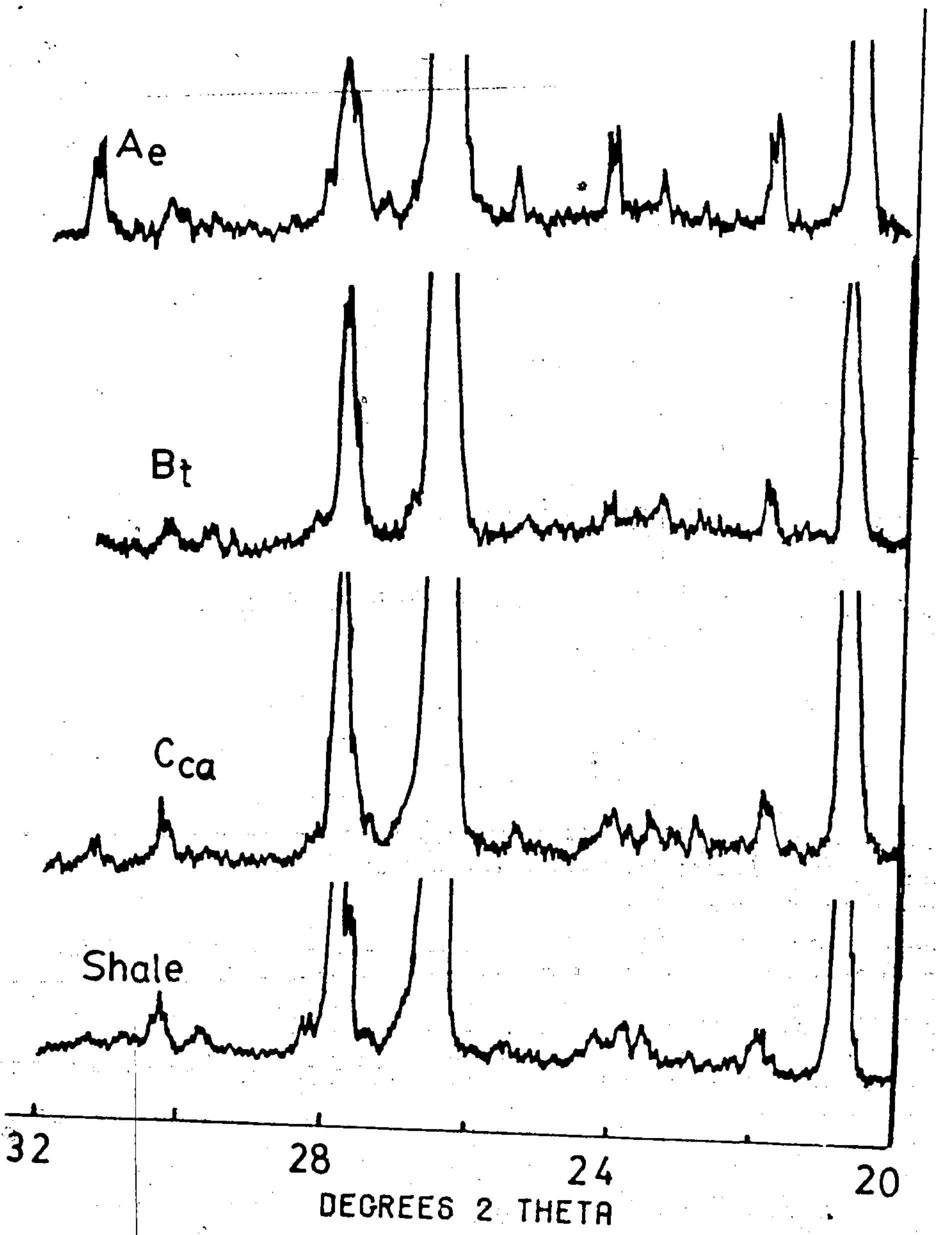


Figure 10. X-ray diffraction patterns for the 2.59 - 2.72 g/cc specific gravity separate from the fine sand fraction of the major genetic horizons from an Orthic Gray Luvisol (site 1)..

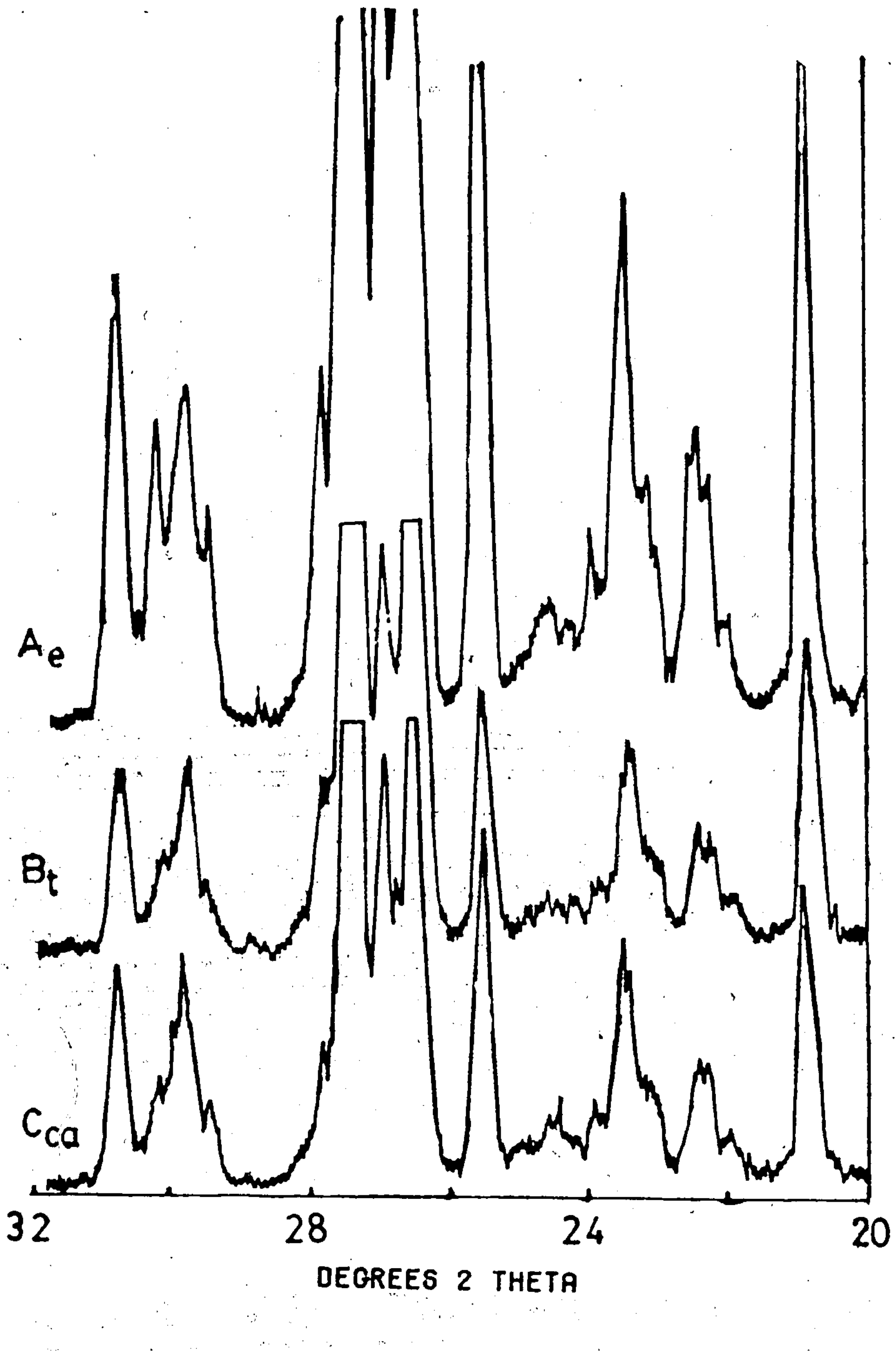


Figure 11. X-ray diffractograms for the 2.53 - 2.59 g/cc specific gravity separates from the coarse and medium silt fractions of the major genetic horizons of an Orthic Gray Luvisol (site 1).

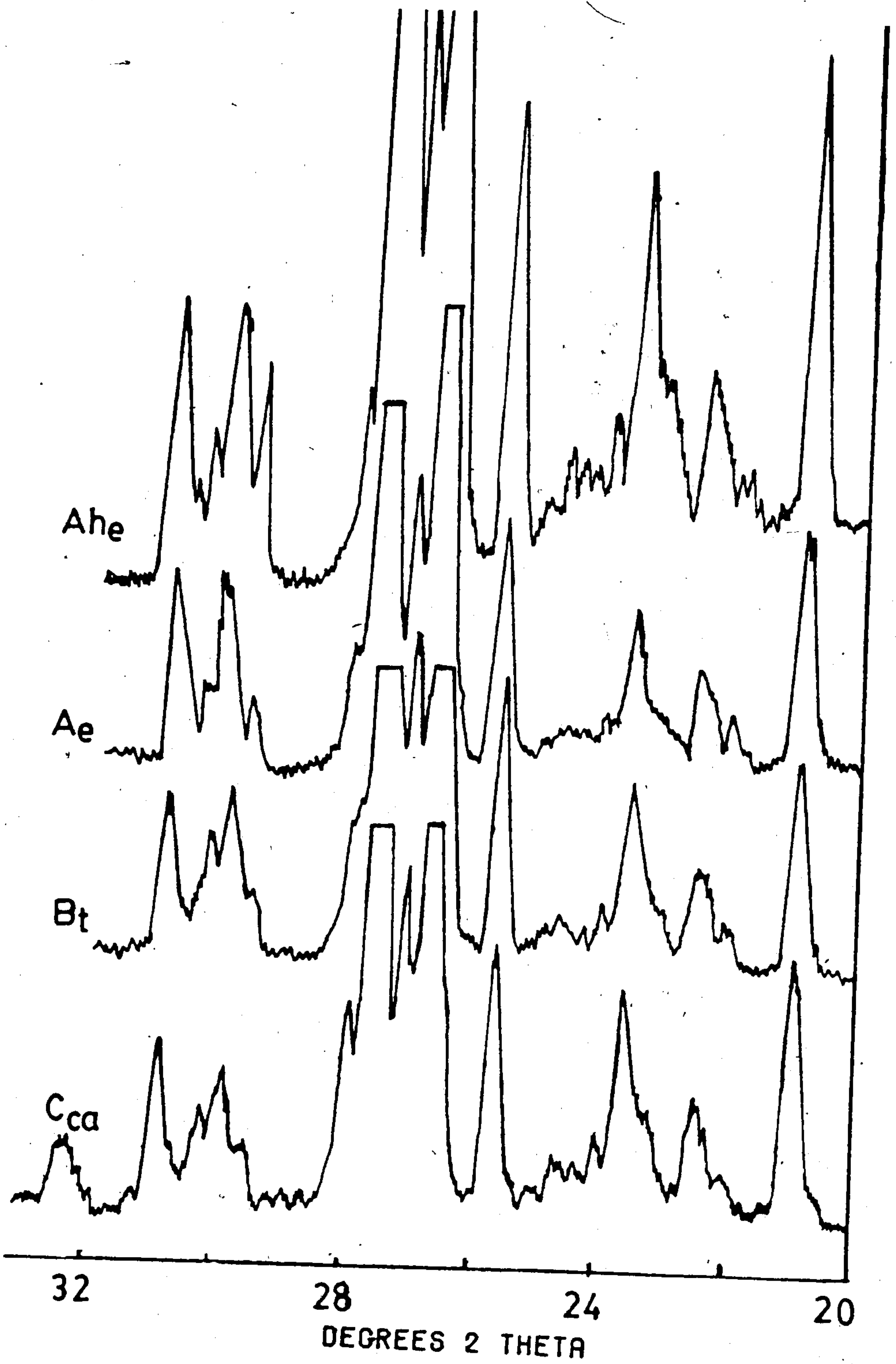


Figure 12. X-ray diffractograms for the 2.53 - 2.59 g/cc specific gravity separates from the coarse and medium silt fractions of the major genetic horizons of a Dark Gray Luvisol (site 2)..............................

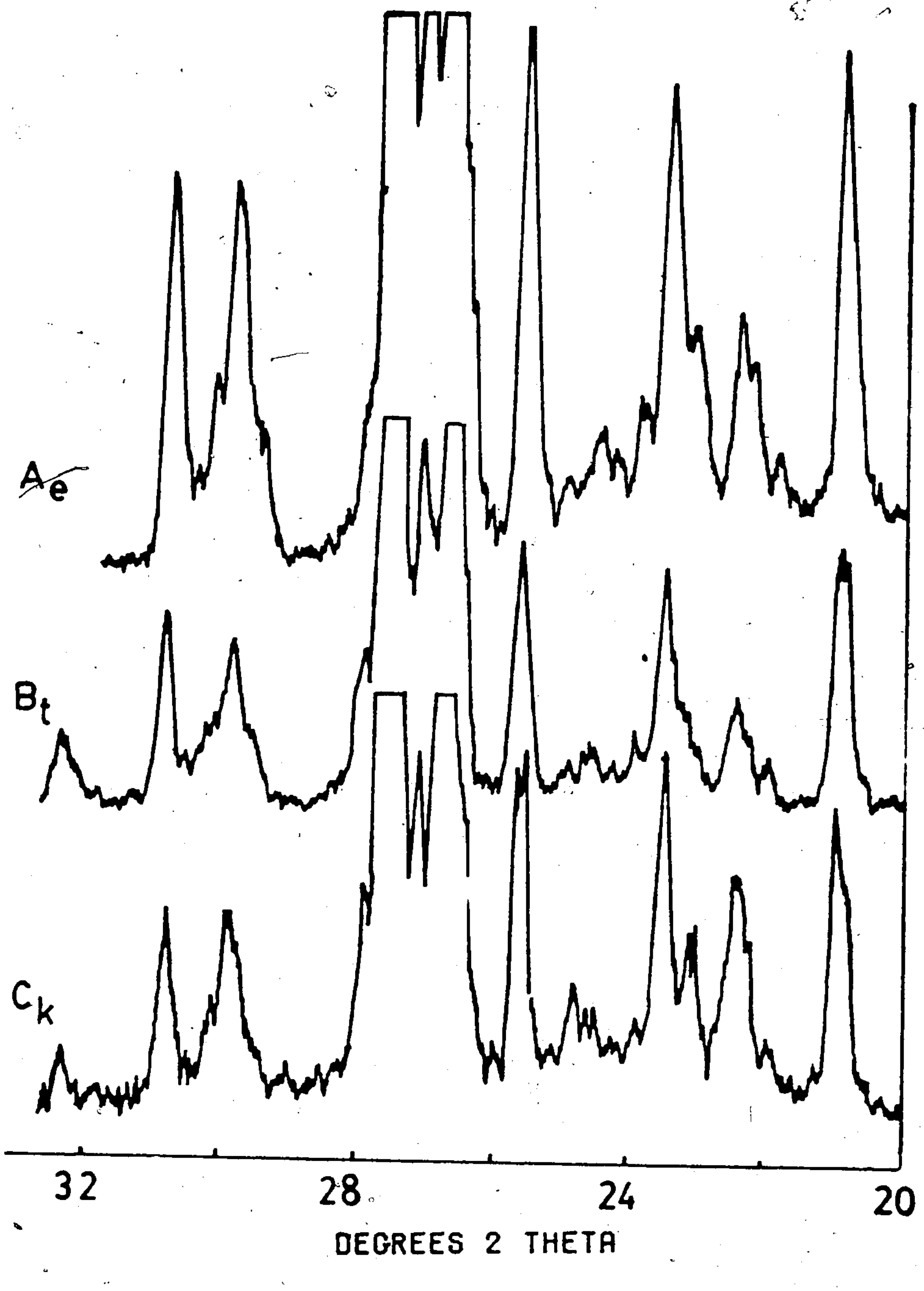


Figure 13. X-ray diffractograms for the 2.53 - 2.59 g/cc specific gravity separates from the coarse and medium silt fractions of the major genetic horizons of an Orthic Gray Luvisol (site 4)..........................

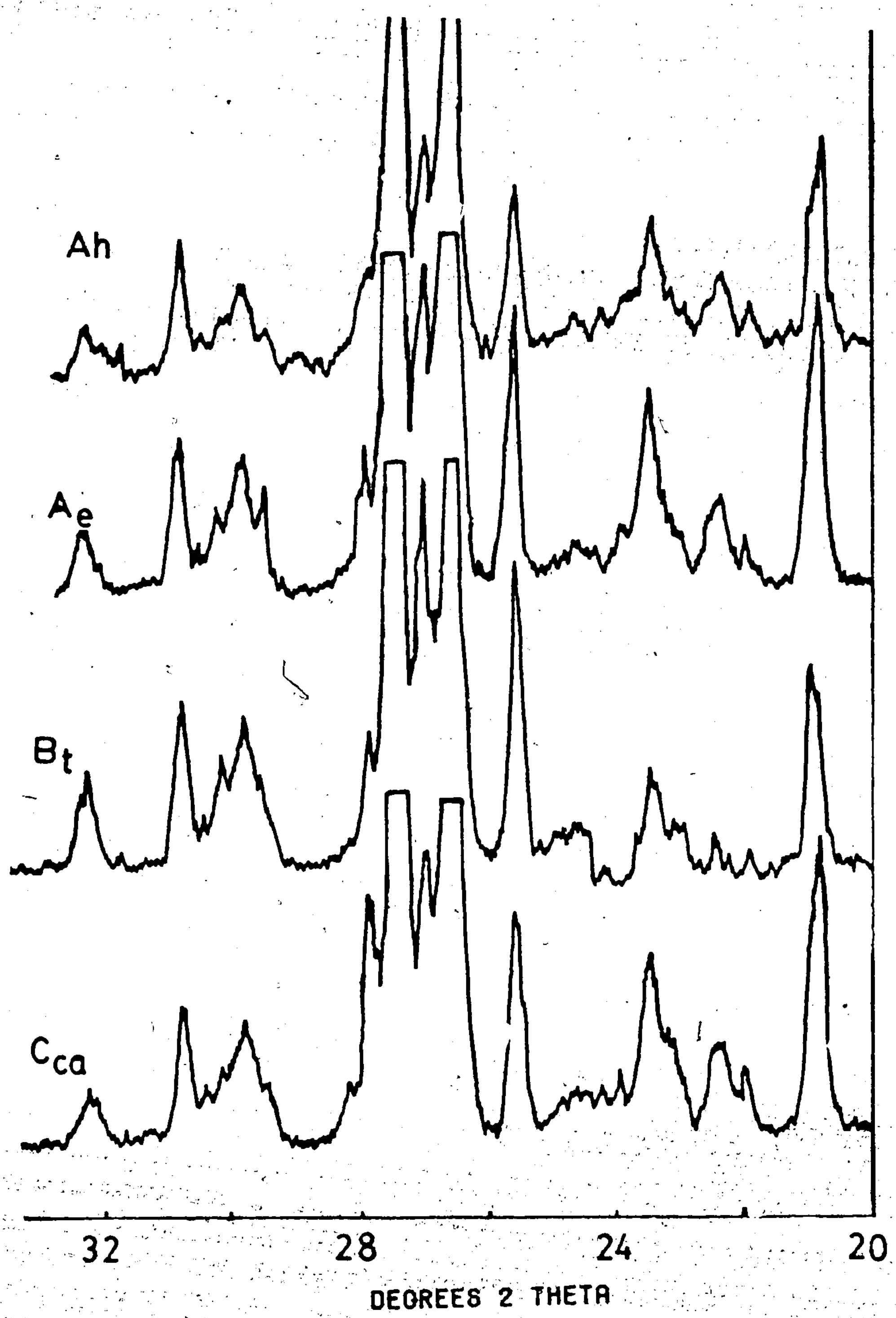


Figure 14. X-ray diffractograms for the 2.53 - 2.59 g/cc specific gravity separates from the coarse and medium silt fractions of the major genetic horizons of the Eluviated Black Chernozem (site 5)........................

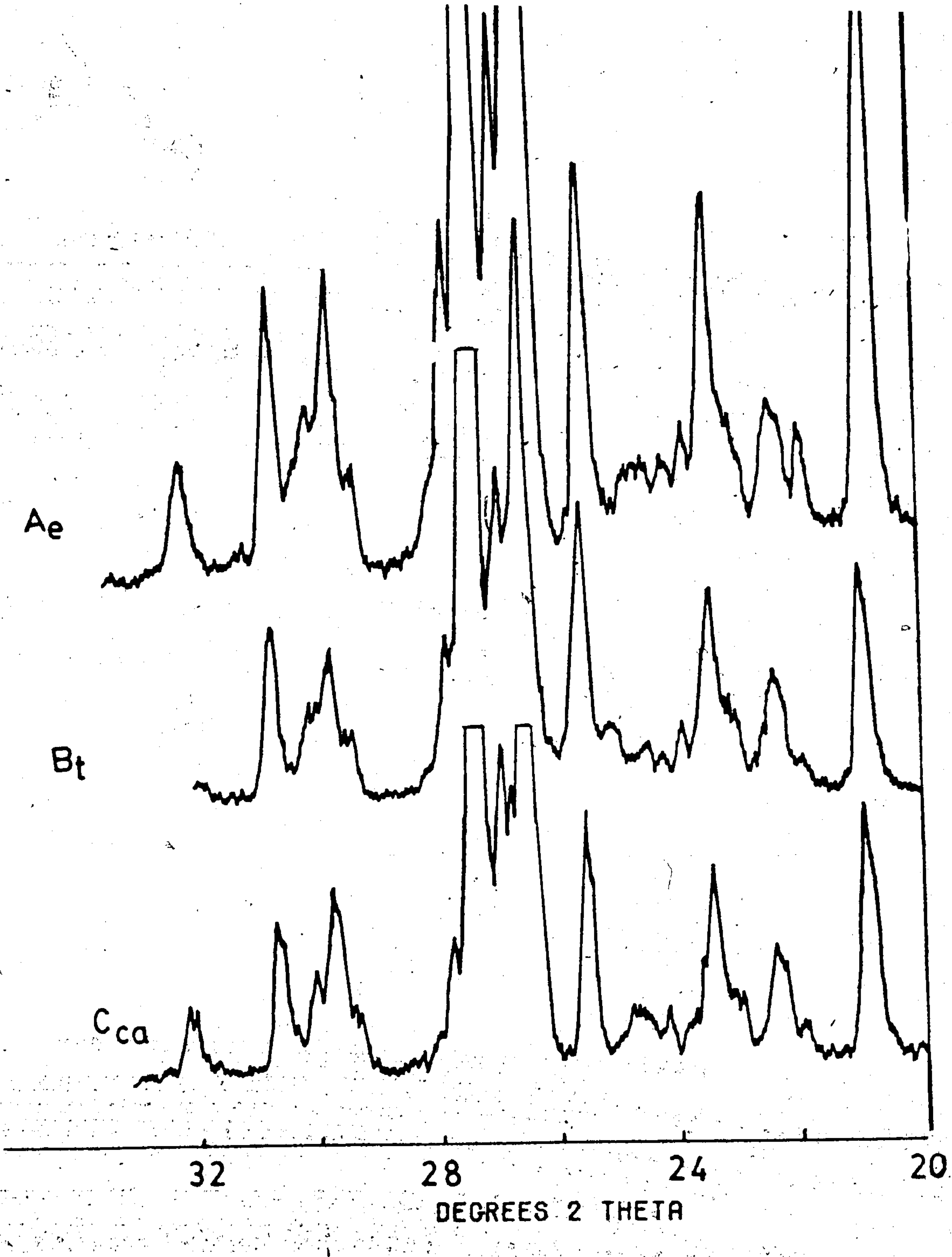


Figure 15. X-ray diffractograms for the 2.53 - 2.59 g/cc specific gravity separates from the coarse and medium silt fractions of the major genetic horizons of an Orthic Gray Luvisol (site 6)..............................

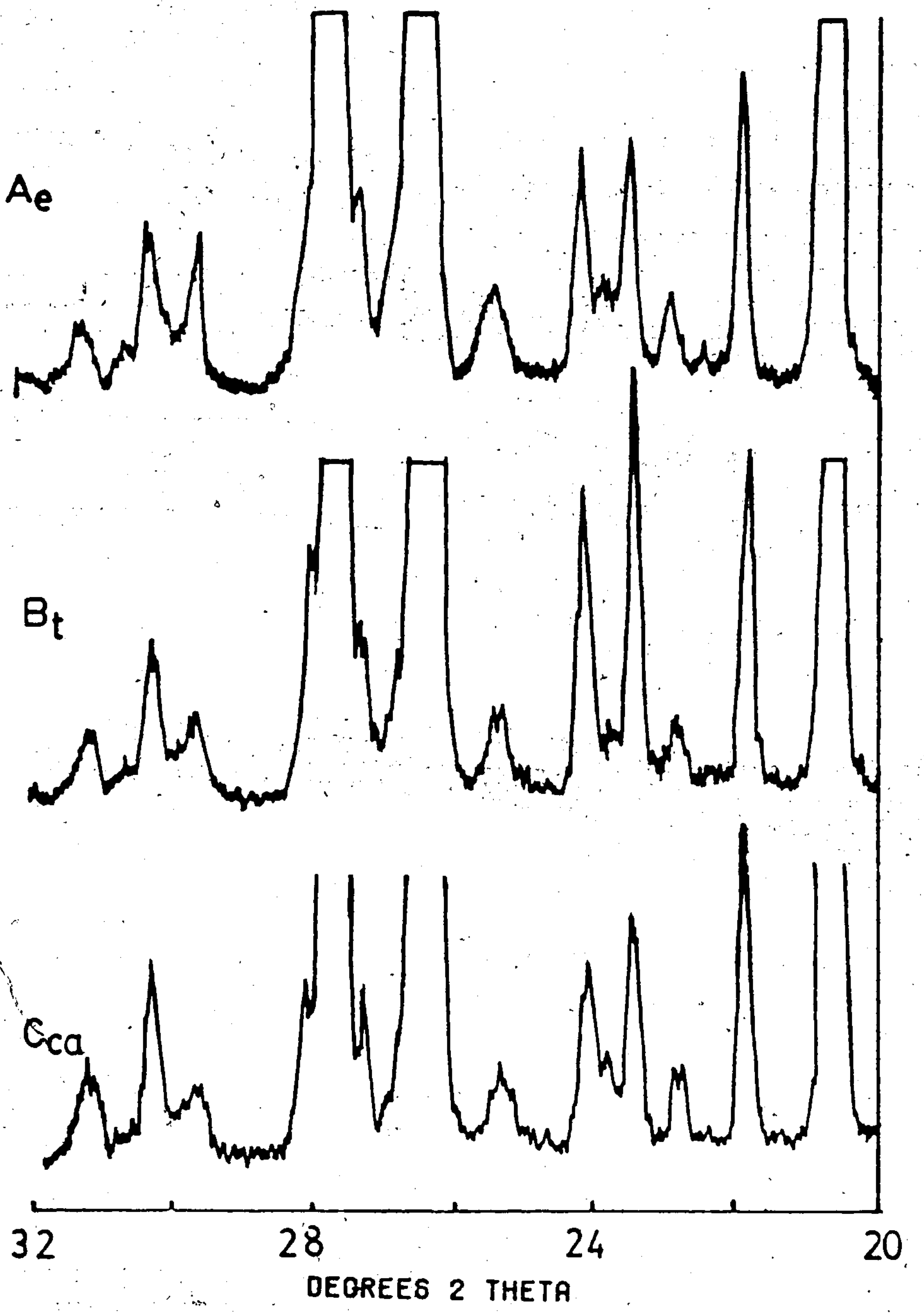


Figure 16. X-ray diffractograms for the 2.59 – 2.72 g/cc specific gravity separates from the coarse and medium silt fractions of the major genetic horizons of an Orthic Gray Luvisol (site 1)..............................

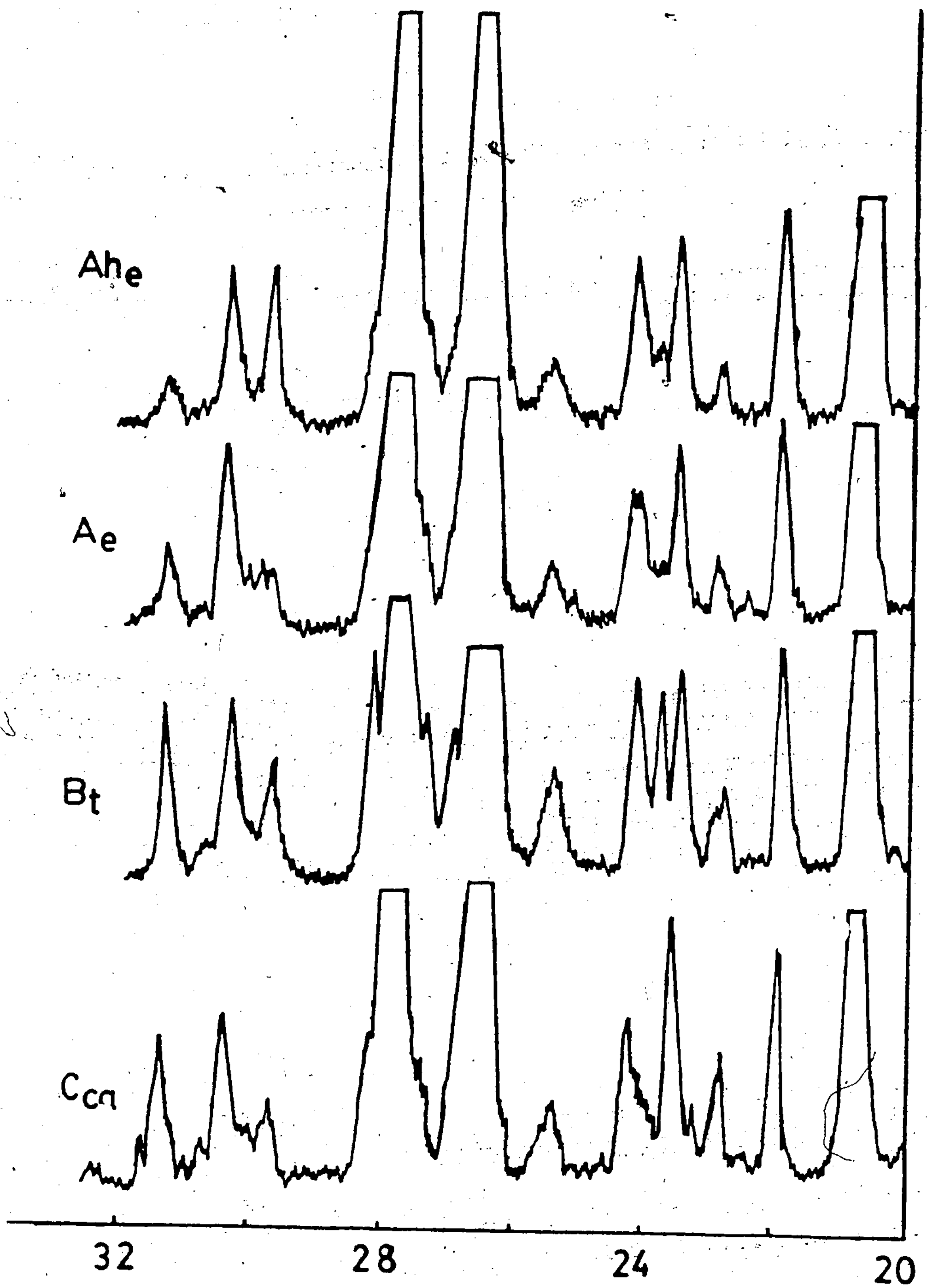


Figure 17. X-ray diffractograms for the 2.59 - 2.72 g/cc specific gravity separates from the coarse and medium silt fractions of the major genetic horizons of a Dark Gray Luvisol (site 2)..............................

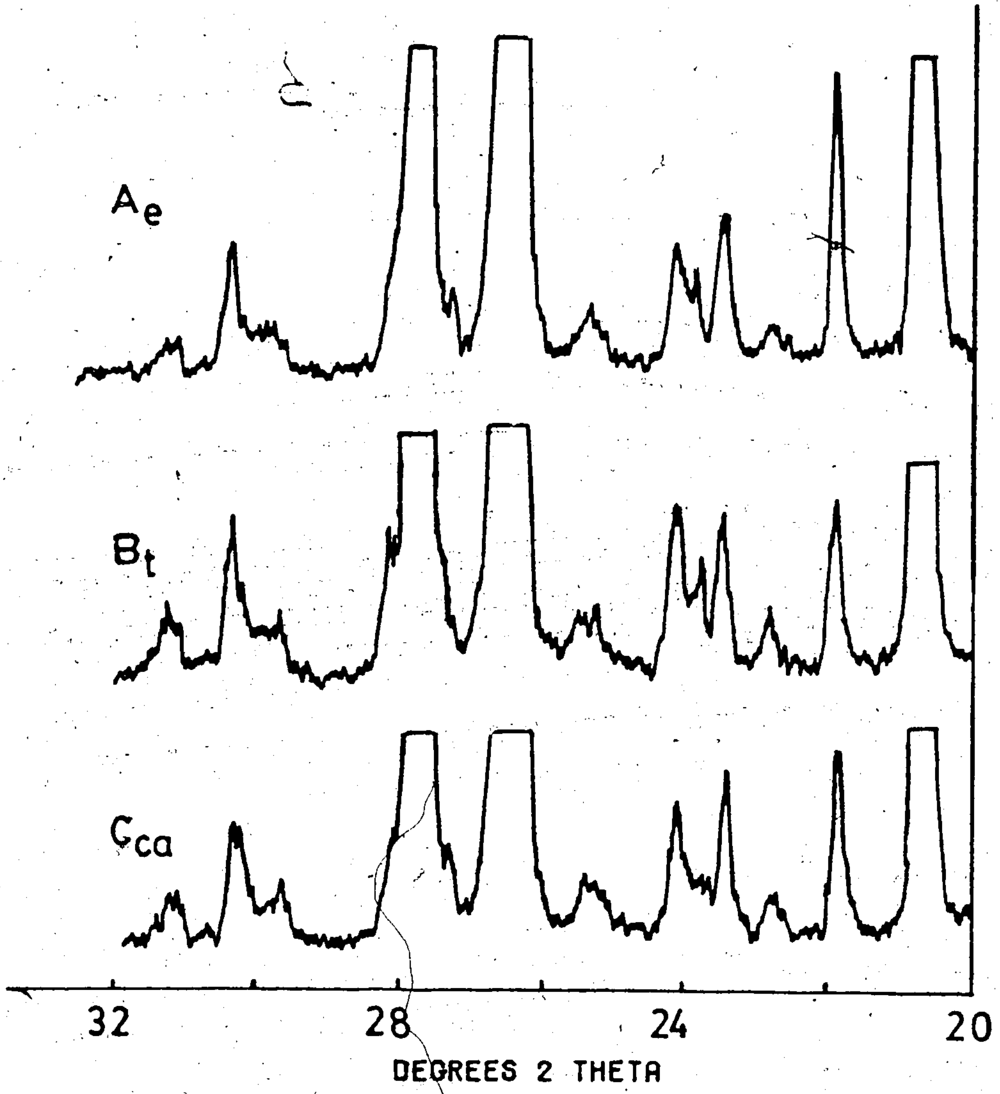


Figure 18. X-ray diffractograms for the 2.59 - 2.72 g/cc specific gravity separates from the coarse and medium silt fractions of the major genetic horizons of an Orthic Gray Luvisol (site 6)..............................

Table 7. Mineralogical composition of the fine sand and coarse and medium silt fractions of the Orthic Gray Luvisol at site 1.

	% (weight basis)				
Horizon	Alkali-feldspars	Plagioclase feldspars	Quartz	Carbonates	Heavy minerals
		Fine sand (50-250 micrometer)			
Ae	6.64	10.9	80.04	nil	2.20
AB	7.04	11.41	79.63	=	1.60
Bt1	6.21	12.63	78.64	=	2.11
Bt2	6.35	11.95	79.05	=	2.18
BT3	6.86	12.72	77.06	0.4	2.76
Cca	7.11	12.57	76.97	1.8	2.73
		Coarse and medium silt (5-50 micrometer)			
Ae	10.66	19.66	64.76	nil	4.92
AB	11.22	18.94	65.17	=	4.67
Bt1	11.19	17.23	66.80	=	4.78
Bt2	11.5	18.23	65.11	=	5.16
BCk	12.79	19.93	61.65	=	5.63
Cca	12.05	19.56	62.32	12.3	6.07

Table 8. Mineralogical composition of the fine sand and coarse and medium silt fractions of the Dark Gray Luvisol at site 2.

	% (weight basis)				
Horizon	Alkali-feldspars	Plagioclase feldspars	Quartz	Carbonates	Heavy minerals
		Fine sand (50-250 micrometer)			
Ahe	6.38	11.65	79.48	nil	2.04
Ae	6.59	10.82	79.94	=	1.89
Bt	6.53	11.34	78.89	=	2.87
Cca1	7.18	12.42	76.77	1.8	2.55
Cca2	6.84	12.19	77.48	4.9	2.04
Shale	12.46	14.18	61.28	ND	11.95
		Coarse and medium silt (5-50 micrometer)			
Ahe	10.80	14.57	68.70	nil	5.93
Ae	10.01	15.10	69.10	=	5.79
Bt	11.33	16.28	65.62	=	6.77
Cca1	10.25	17.23	64.73	14.3	7.79
Cca2	12.31	18.30	62.68	13.0	6.71
Shale	10.61	18.23	62.81	ND	8.35

Table 9. Mineralogical composition of the fine sand and coarse and medium silt fractions of the Dark Gray Luvisol at site 3.

	% (weight basis)				
Horizon	Alkali-feldspars	Plagioclase feldspars	Quartz	Carbonates	Heavy minerals
		<u>Fine sand (50-250 micrometer)</u>			
Ahe	7.52	12.13	77.72	nil	2.40
Ae	7.65	12.43	76.83	=	2.47
AB	8.09	12.66	76.61	=	2.09
Bt1	7.61	13.85	75.88	=	2.53
Bt2	7.89	11.98	77.67	=	2.30
BCk	7.18	12.12	76.70	0.4	2.30
Cca	7.20	12.65	76.42	1.3	2.75
		<u>Coarse and medium silt (5-50 micrometer)</u>			
Ahe	10.73	14.83	69.62	nil	4.82
Ae	10.82	15.11	68.36	=	5.71
AB	11.39	16.86	66.09	=	5.66
Bt1	11.65	16.02	66.29	=	6.04
Bt2	12.61	18.15	63.73	=	5.51
BCk	12.06	17.99	62.82	=	7.13
Cca	12.08	18.51	62.50	11.9	6.91

Table 10. Mineralogical composition of the fine sand and coarse and medium silt fractions of the Orhic Gray Luvisol at site 4.

Horizon	% (weight basis)				
Horizon	Alkali-feldspars	Plagioclase feldspars	Quartz	Carbonates	Heavy minerals
		Fine sand (50-250 micrometer)			
Ae	5.91	7.93	83.96	nil	1.59
AB	6.07	8.26	83.87	=	1.54
Bt1	6.46	9.89	81.55	=	1.55
Bt2	6.51	11.02	80.02	=	1.82
Ck	6.13	11.46	78.81	1.8	2.56
		Coarse and medium silt (5-50 micrometer)			
Ae	8.97	11.51	74.40	nil	5.12
AB	6.69	10.05	77.75	=	5.51
Bt1	10.56	11.00	71.97	=	6.97
Bt2	10.21	12.18	69.05	=	8.56
Ck	9.28	13.08	69.49	=	8.15

Table 11. Mineralogical composition of the fine sand and coarse and medium silt fractions of the Eluviated Black Chernozemic at site 5.

	% (weight basis)				
Horizon	Alkali-feldspars	Plagioclase feldspars	Quartz	Carbonates	Heavy minerals
		Fine sand (50-250 micrometer)			
Ahe	6.31	10.71	79.97	nil	2.38
Ae	5.87	9.34	82.65	=	1.93
Bt1	5.95	10.02	81.17	=	2.36
Bt2	5.33	10.12	81.89	=	2.17
BCk	6.44	10.09	80.66	1.6	2.16
Cca	6.26	11.45	79.08	2.2	2.34
		Coarse and medium silt (5-50 micrometer)			
Ahe	10.58	14.09	71.47	nil	3.86
Ae	10.66	13.70	71.74	=	3.90
Bt1	10.48	14.62	69.77	=	5.13
Bt2	10.49	15.89	67.85	=	5.77
BCk	10.74	16.63	66.74	12.8	5.89
Cca	10.41	18.11	65.98	14.9	5.50

Table 12. Mineralogical composition of the fine sand and coarse and medium silt fractions of the Orthic Gray Luvisol at site 6.

	% (weight basis)				
Horizon	Alkali- feldspars	Plagioclase feldspars	Quartz	Carbonates	Heavy minerals
		Fine sand (50-250 micrometer)			
Ae	4.38	7.36	86.63	nil	1.23
AB	4.76	7.11	86.22	=	1.55
Bt	3.95	6.41	87.65	=	1.37
BCk	4.40	7.36	86.42	1.0	1.32
Cca	4.21	6.65	87.31	2.7	1.56
		Coarse and medium silt (5-50 micrometer)			
Ae	6.85	13.39	76.51	nil	3.25
AB	9.09	13.86	72.62	=	4.43
Bt	8.53	12.46	74.54	=	4.47
BCk	9.06	12.35	73.64	2.8	4.95
Cca	8.97	12.73	73.45	15.6	4.85

Table 13. Relative feldspar composition of the 2.53 to 2.59 g/cc and 2.59 to 2.72 g/cc sand-sized specific gravity separates as determined from bulk chemical composition.

Soil	Horizon	Relative feldspar composition (Mole%)							
		2.53 to 2.59 g/cc			2.59 to 2.72 g/cc				
					Determined			Recalculated*	
		Ab.	Or.	An.	Ab.	Or.	An.	Ab.	An.
Orthic Gray Luvisol (Site 1)	Ae	15	84	1	67	13	20	77	23
	AB	14	85	1	68	12	20	77	23
	Bt1	17	82	1	66	14	20	77	23
	Bt2	19	79	2	63	17	20	76	24
	BCk	18	81	1	67	14	19	78	22
	Cca	18	81	1	65	14	21	76	24
Dark Gray Luvisol (Site 2)	Ahe	17	82	1	61	23	16	79	21
	Ae	16	83	1	63	20	17	79	21
	Bt	14	85	1	62	17	21	75	25
	Cca1	18	81	1	68	13	19	78	22
	Cca2	19	79	2	68	13	19	78	22
Dark Gray Luvisol (Site 3)	Ahe	14	85	1	62	24	14	82	18
	Ae	15	84	1	59	25	16	79	21
	AB	14	85	1	58	26	16	78	22
	Bt1	15	84	1	62	22	16	80	20
	Bt2	16	83	1	58	25	17	77	23
	BCk	21	78	1	65	15	20	77	23
	Cca	17	82	1	67	14	19	80	20
Orthic Gray Luvisol (Site 4)	Ae	15	84	1	58	30	12	83	17
	AB	14	85	1	58	32	10	85	15
	Bt1	15	83	2	59	27	14	81	19
	Bt2	20	77	3	61	23	16	79	21
	Ck	18	80	2	63	19	18	78	22
Eluviated Black Chernozemic (site 5)	Ah	13	86	1	61	22	17	78	22
	Ae	12	87	1	59	25	16	79	21
	Bt1	13	86	1	60	22	18	77	23
	Bt2	13	86	1	63	19	18	78	22
	BCk	13	86	1	58	25	17	77	23
	Cca	15	84	1	62	20	18	76	24
Orthic Gray Luvisol (Site 6)	Ae	14	85	1	64	25	11	85	14
	AB	16	82	2	57	34	9	86	14
	Bt	14	85	1	59	25	16	79	21
	BCk	15	84	1	59	25	16	79	21
	Cca	15	84	1	58	26	16	78	22
Shale		17	80	3	59	19	22	73	27

*recalculated to an orthoclase free basis

Table 14. Relative feldspar composition of the 2.53 to 2.59 g/cc and 2.59 to 2.72 g/cc silt-sized specific gravity separates as determined from bulk chemical composition.

Soil	Horizon	Relative feldspar composition (Mole%)							
		2.53 to 2.59 g/cc			2.59 to 2.72 g/cc				
					Determined			Recalculated*	
		Ab.	Or.	An.	Ab.	Or.	An.	Ab.	An.
Orthic Gray Luvisol (Site 1)	Ae	15	84	1	63	21	16	80	20
	AB	11	88	1	73	13	14	84	16
	Bt1	15	84	1	59	24	17	78	22
	Bt2	14	85	1	58	23	17	77	23
	BCk	15	83	2	61	24	15	80	20
	Cca	15	83	2	68	16	16	81	19
Dark Gray Luvisol (Site 2)	Ahe	13	85	2	60	27	13	82	18
	Ae	13	85	2	69	16	16	81	19
	Bt	11	88	1	63	21	16	80	20
	Cca1	14	85	1	72	10	18	80	20
	Cca2	14	85	1	68	15	17	80	20
Dark Gray Luvisol (Site 3)	Ahe	13	83	4	66	20	14	83	17
	Ae	13	86	1	63	24	13	83	17
	AB	12	87	1	66	19	15	82	18
	Bt1	14	85	1	64	20	16	80	20
	Bt2	13	86	1	64	20	16	80	20
	BCk	12	87	1	72	9	19	79	21
	Cca	12	87	1	65	19	16	77	23
Orthic Gray Luvisol (Site 4)	Ae	10	89	1	62	30	8	89	11
	AB	11	88	1	65	27	8	89	11
	Bt1	11	88	1	60	33	7	90	10
	Bt2	12	87	1	53	27	10	84	16
	Ck	15	83	2	66	23	11	86	14
Eluviated Black Chernozemic (site 5)	Ah	12	81	7	67	19	14	83	17
	Ae	11	88	1	63	24	13	83	17
	Bt1	10	89	1	66	17	16	81	19
	Bt2	15	84	1	67	15	18	79	21
	BCk	20	78	2	68	15	17	80	20
	Cca	16	82	2	70	13	18	80	20
Orthic Gray Luvisol (Site 6)	Ae	14	85	1	75	13	12	86	14
	AB	12	86	2	58	31	10	85	15
	Bt	12	87	1	67	17	16	81	19
	BCk	12	87	1	66	20	14	82	18
	Cca	12	87	1	66	19	15	82	18
Shale		12	86	2	53	22	25	68	32

*recalculated to an orthoclase-free basis

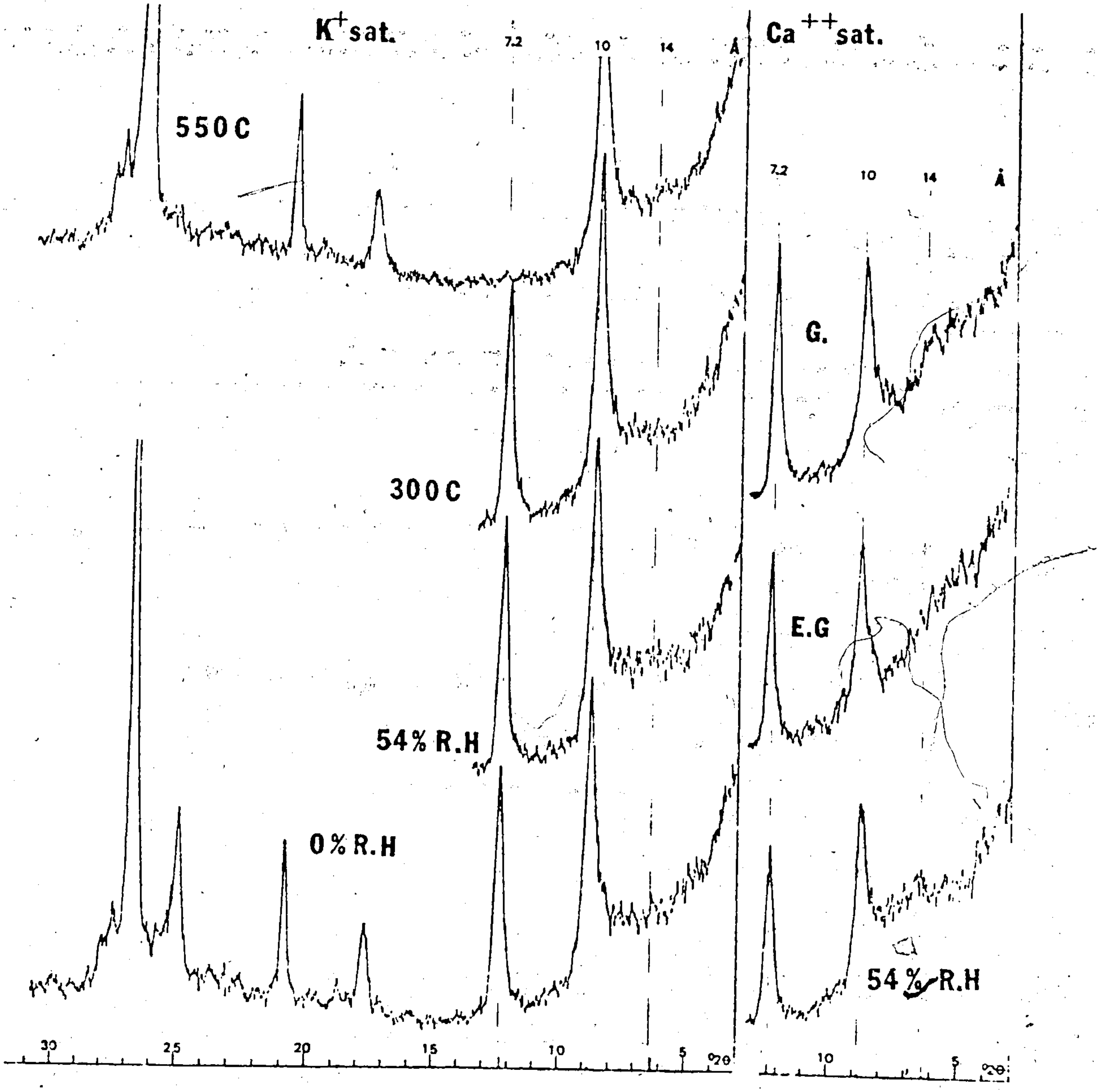


Figure 19. X-ray diffraction patterns of total clay separated from the Ahe horizon of a Dark Gray Luvisol (site 3)

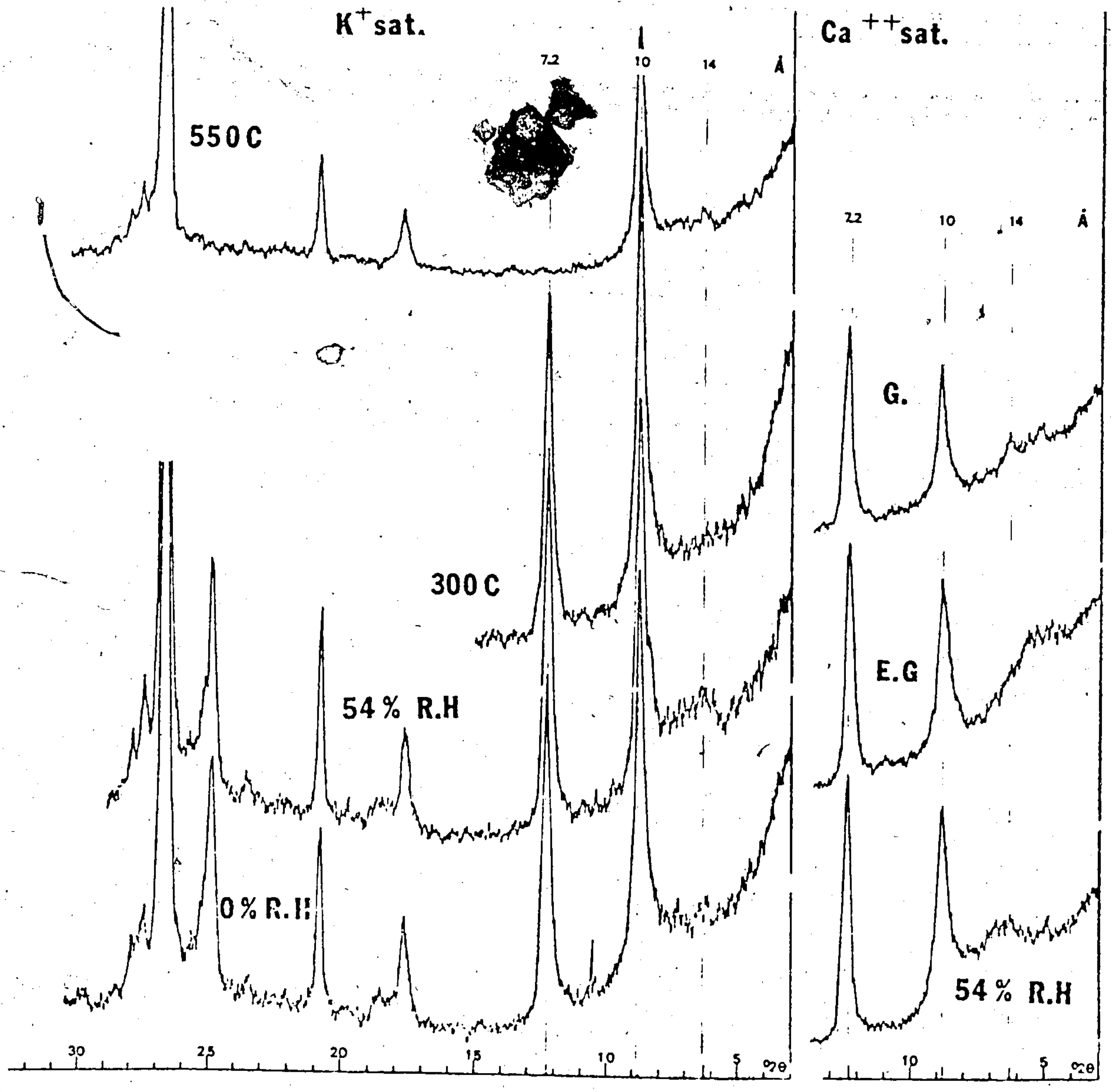


Figure 20. X-ray diffraction patterns of total clay separated from the Ae horizon of a Dark Gray Luvisol (site 3)...............................

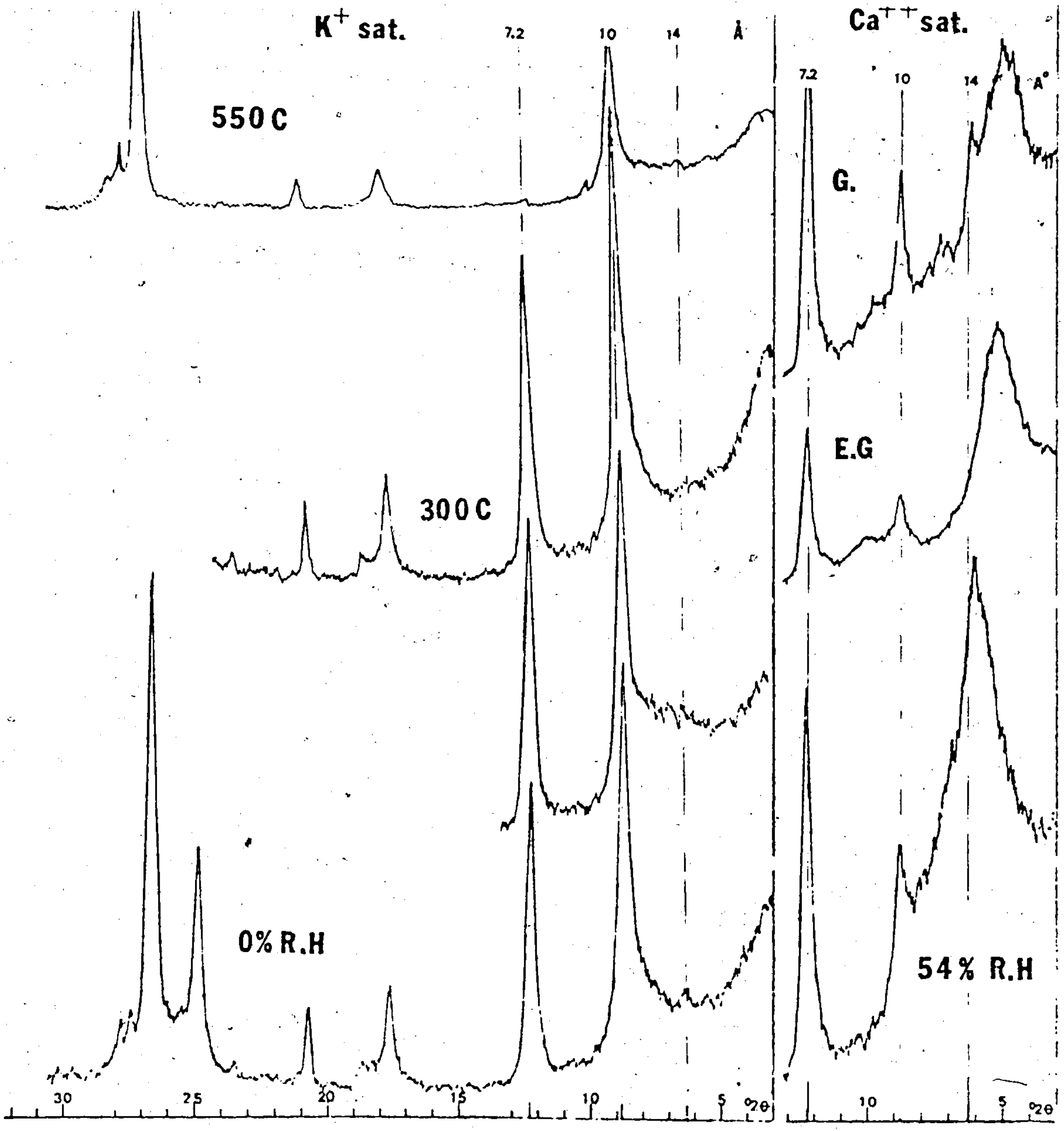


Figure 21. X-ray diffraction patterns of total clay separated from the Bt2 horizon of a Dark Gray Luvisol (site 3)........................

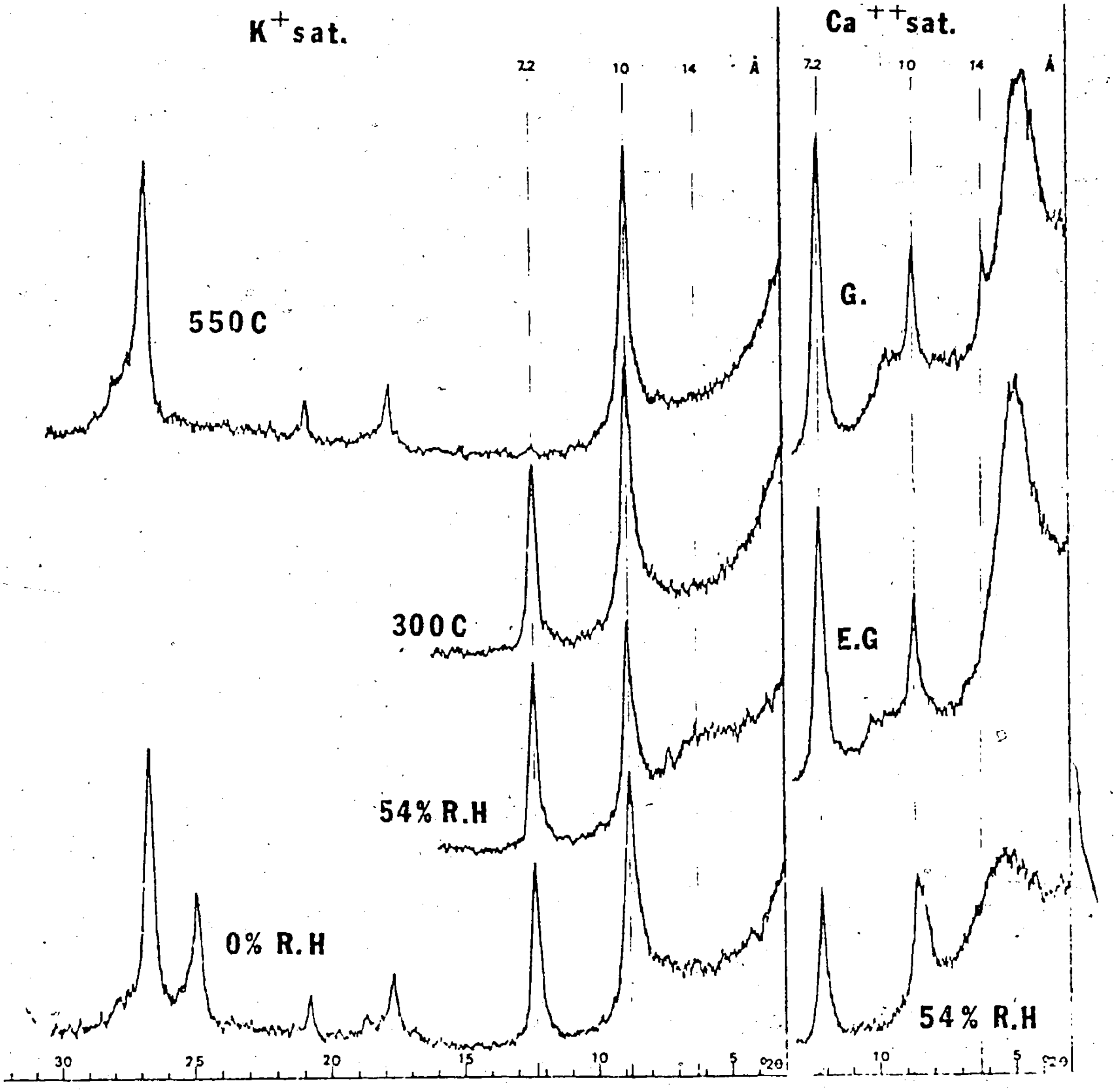


Figure 22. X-ray diffraction patterns of total clay separated from the Cca horizon of a Dark Gray Luvisol (site 3)

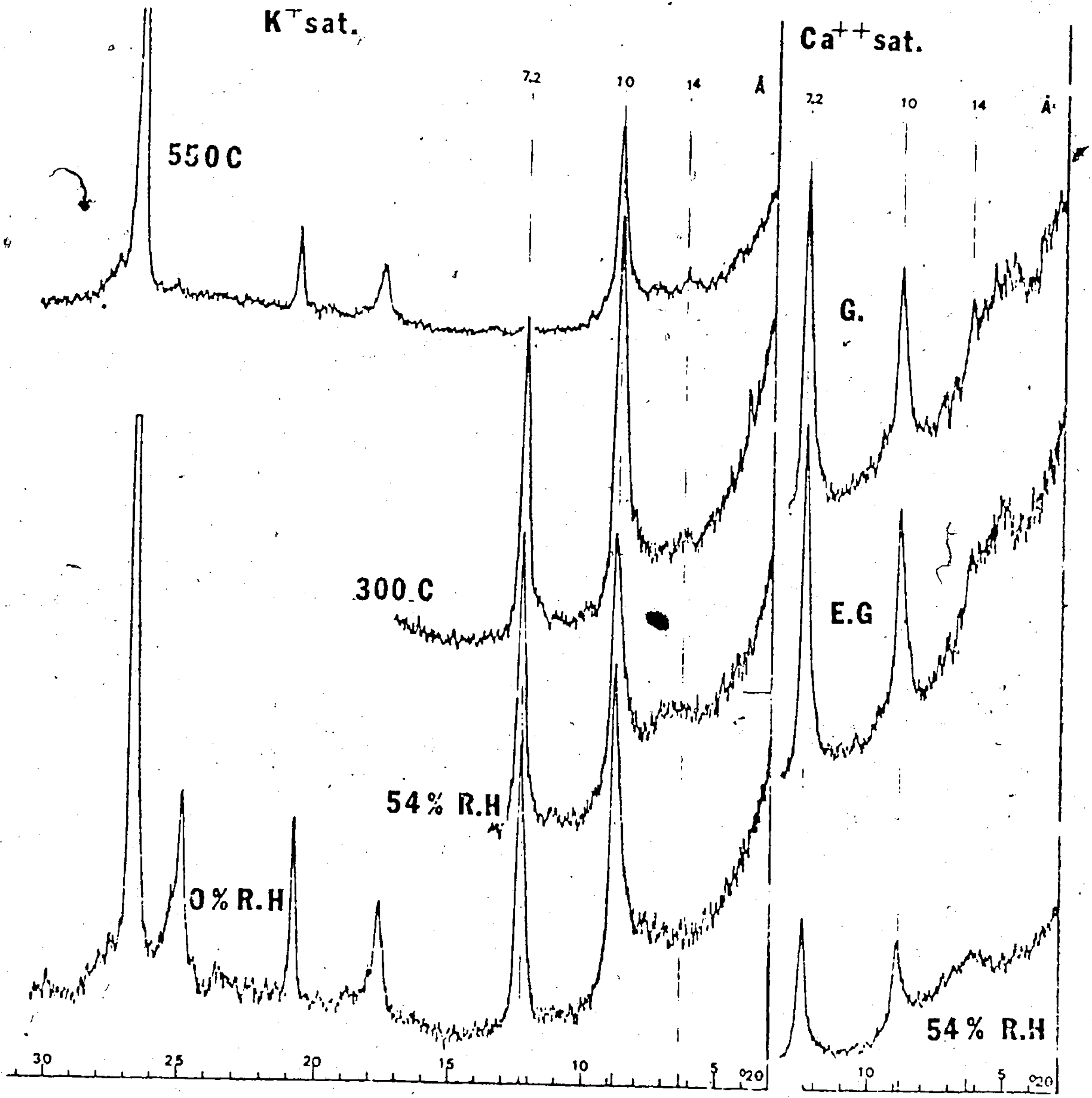


Figure 23. X-ray diffraction patterns of total clay separated from the Ae horizon of an Orthic Gray Luvisol (site 4).

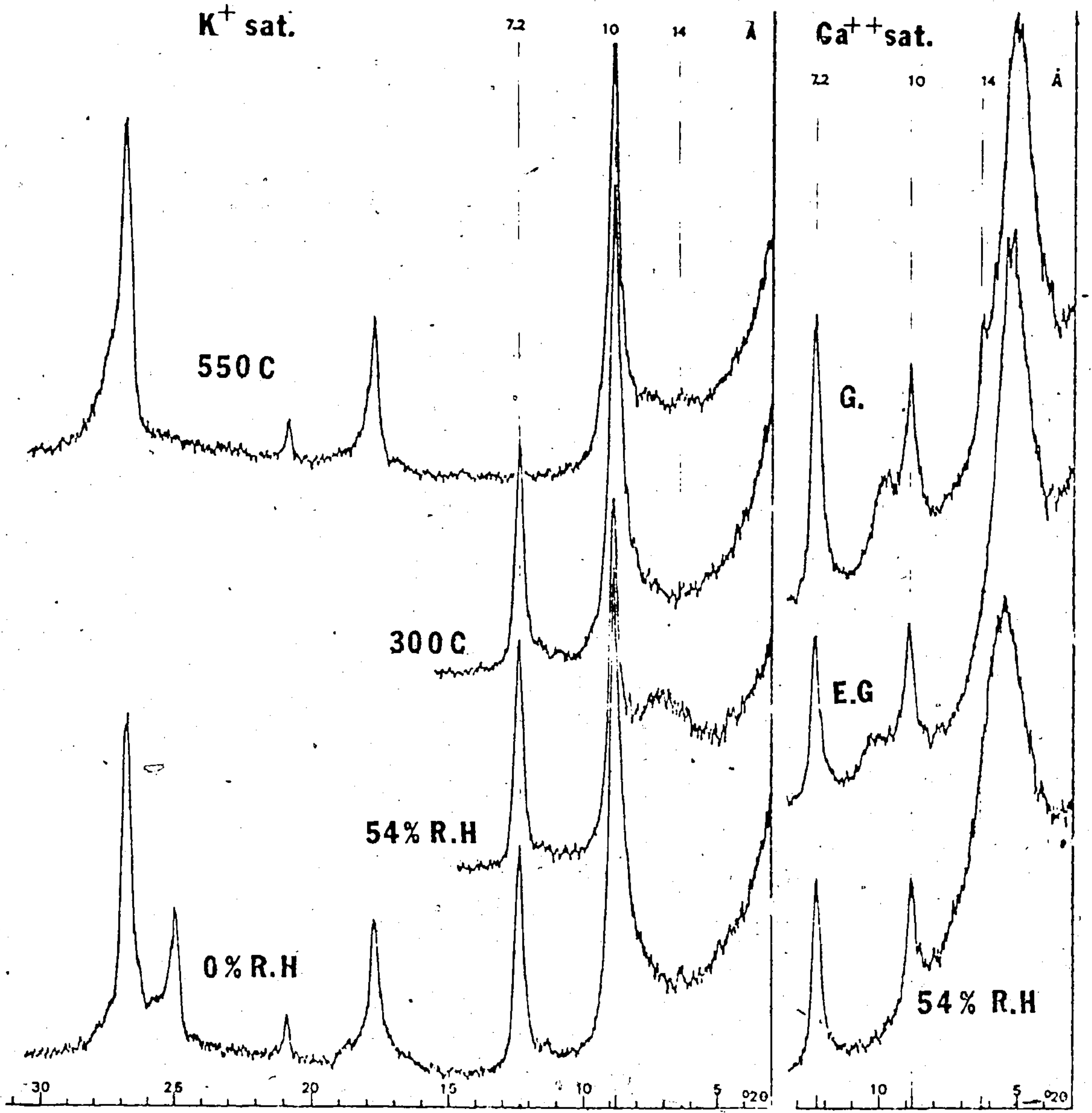


Figure 24. X-ray diffraction patterns of total clay separated from the Bt2 horizon of an Orthic Gray Luvisol (site 4)........................

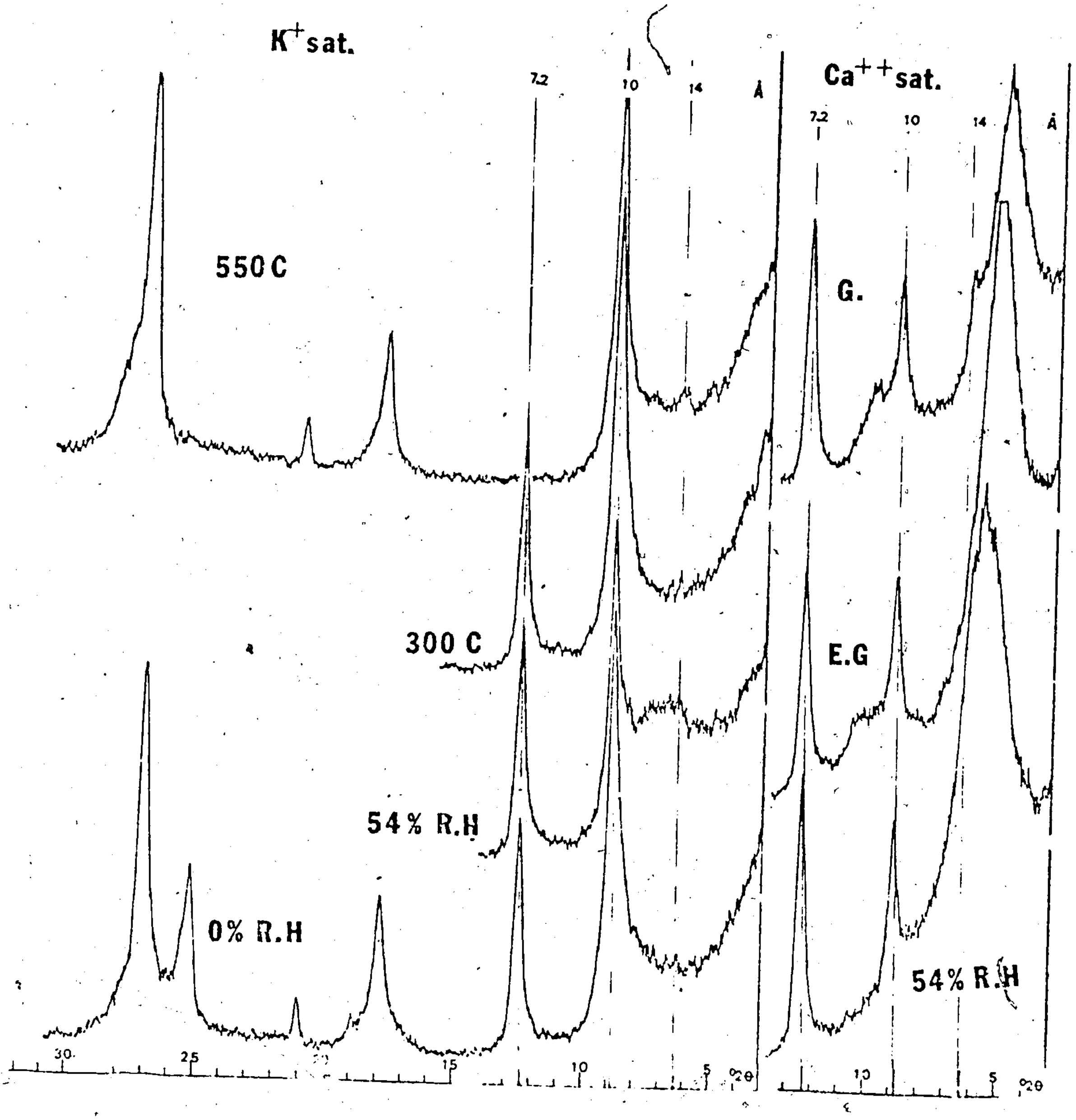


Figure 25. X-ray diffraction patterns of total clay from the Ck horizon of an Orthic Gray Luvisol (site 4)...........................

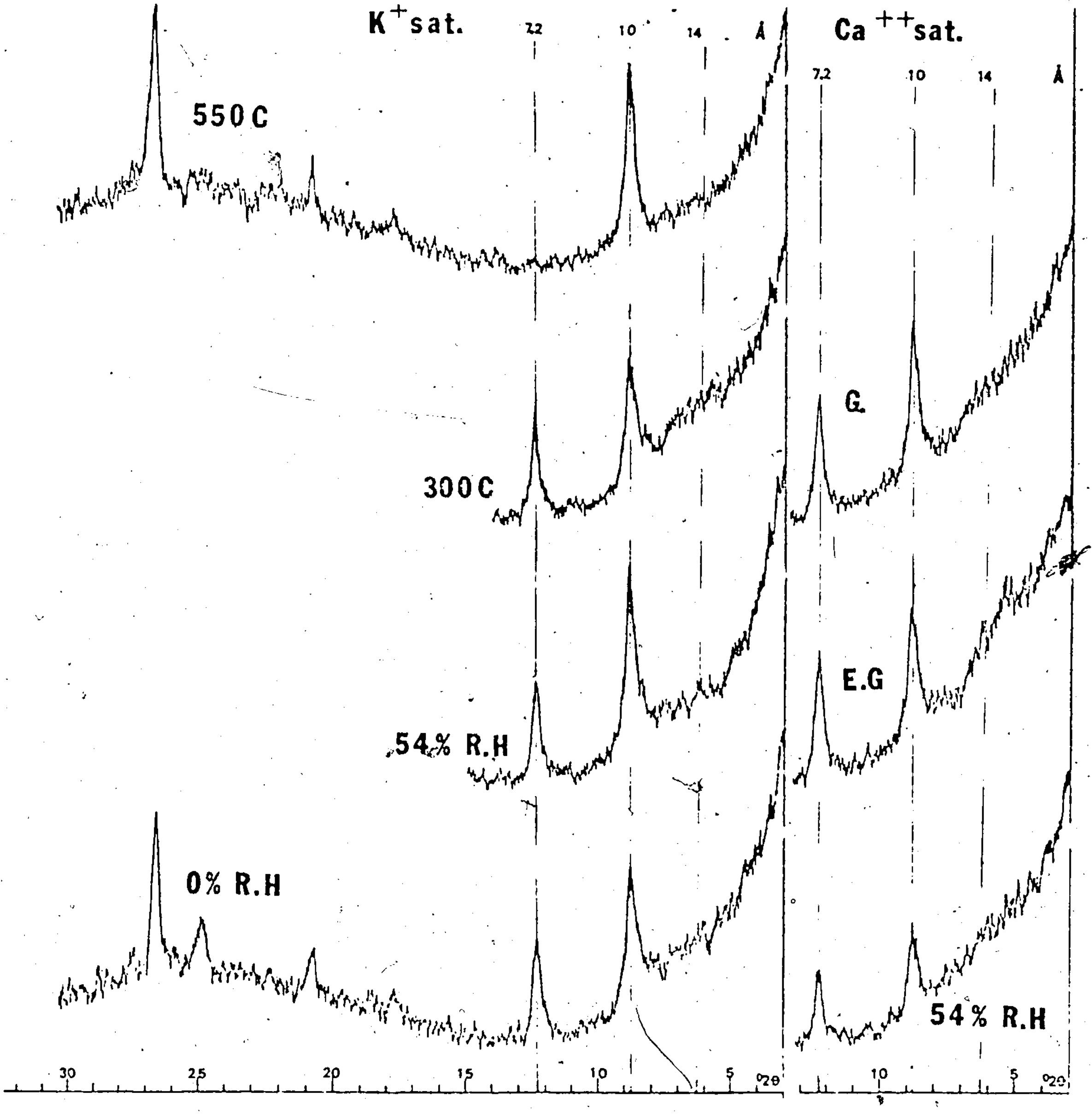


Figure 26. X-ray diffraction patterns of total clay from the Ah horizon of the Eluviated Black Chernozem (site 5)............................

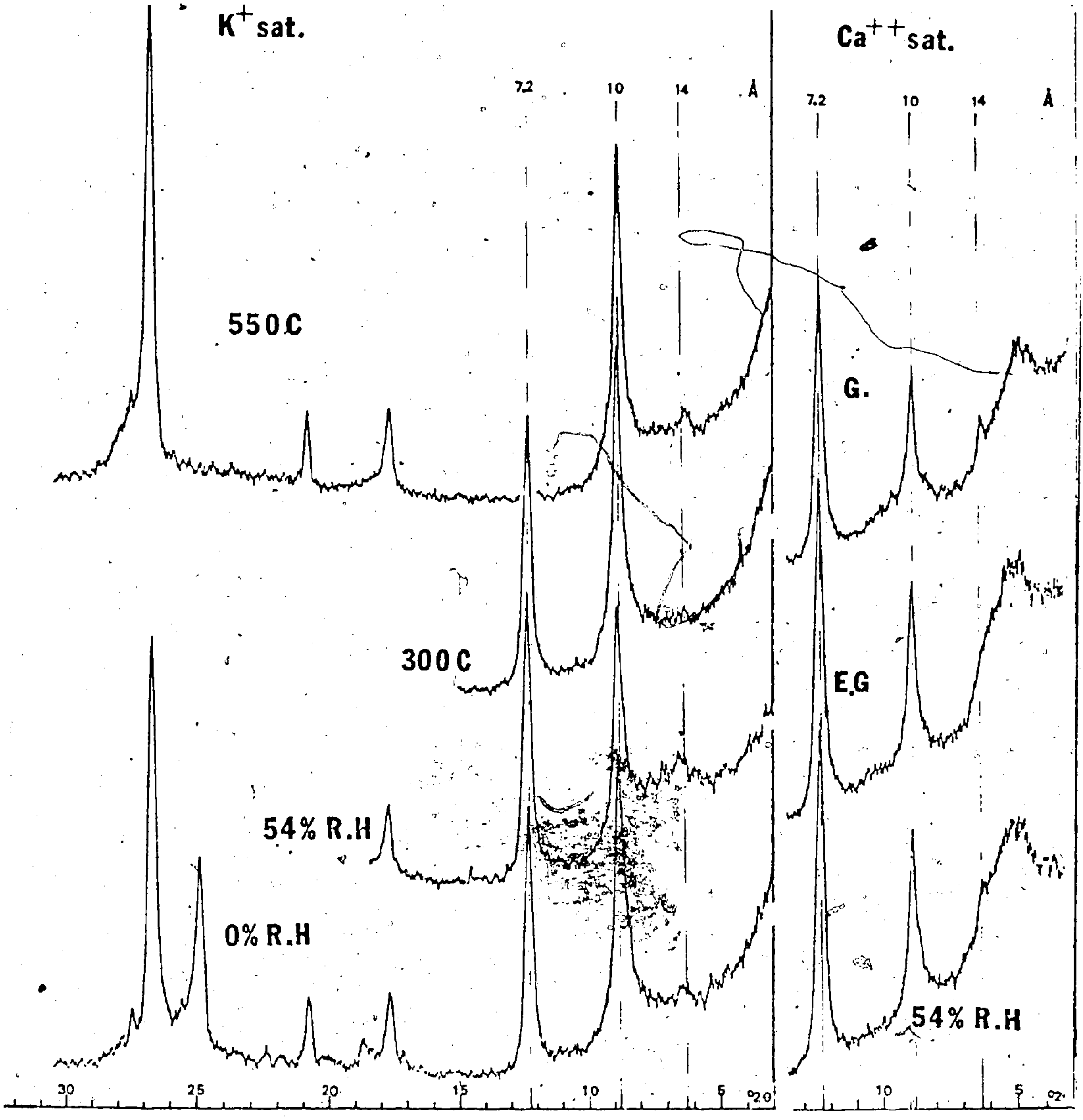


Figure 27. X-ray diffraction patterns of total clay from the Ae horizon of the Eluviated Black Chernozem (site 5)........................ ....

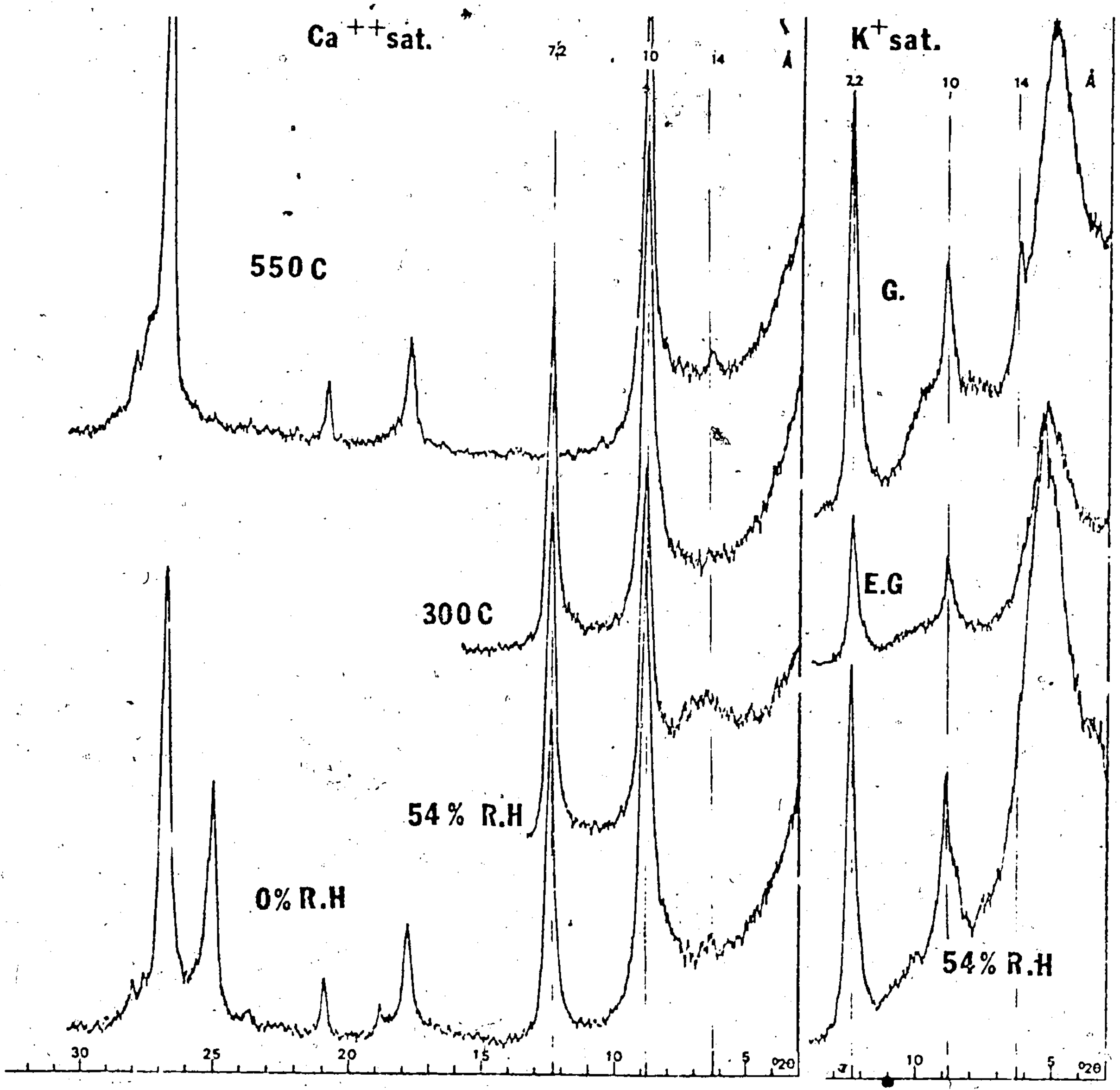


Figure 28. X-ray diffraction patterns of total clay from the Bt horizon of the Eluviated Black Chernozem (site 5)

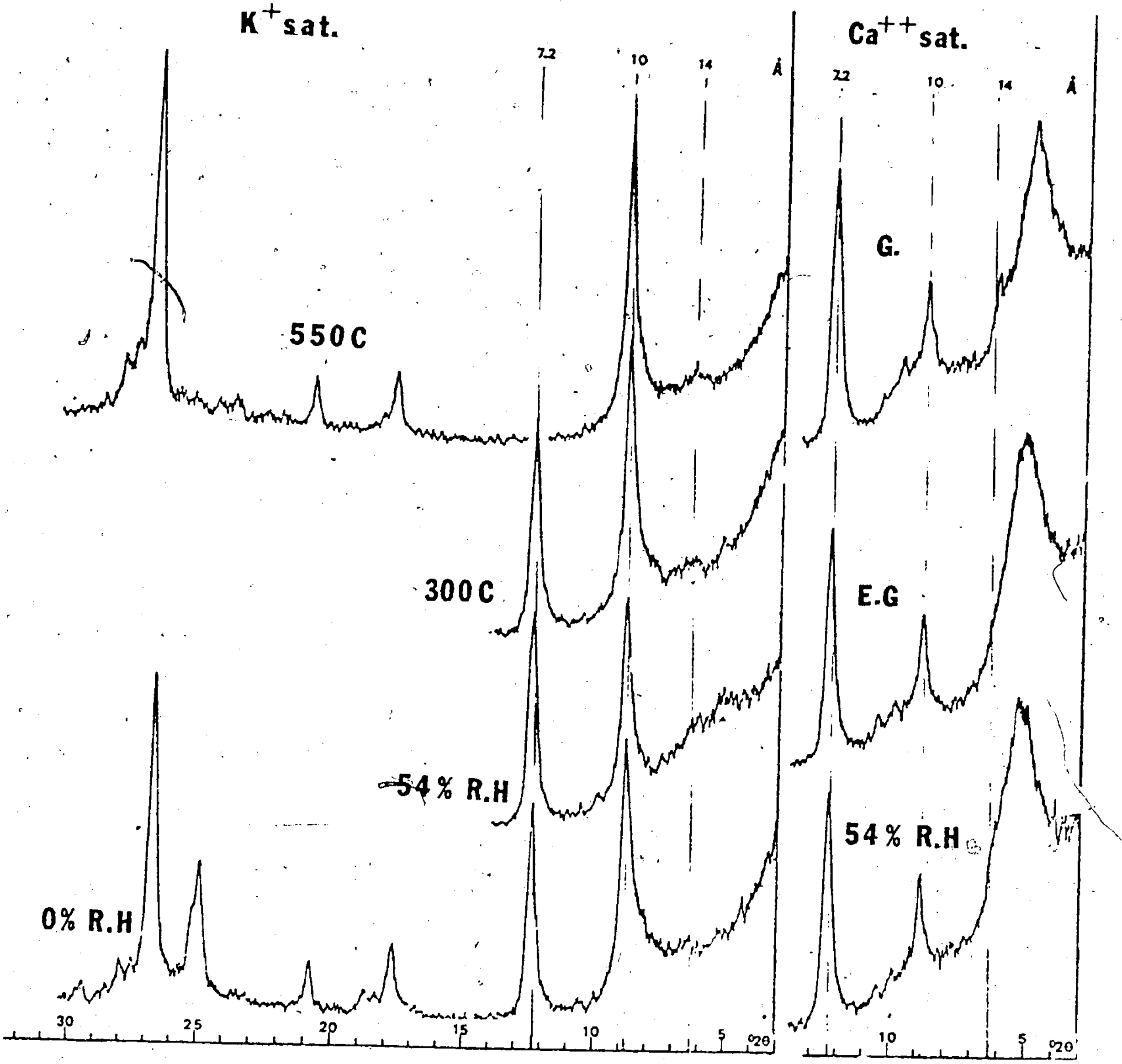


Figure 29. X-ray diffraction patterns of total clay from the Cca horizon of the Eluviated Black Chernozem (site 5)............................

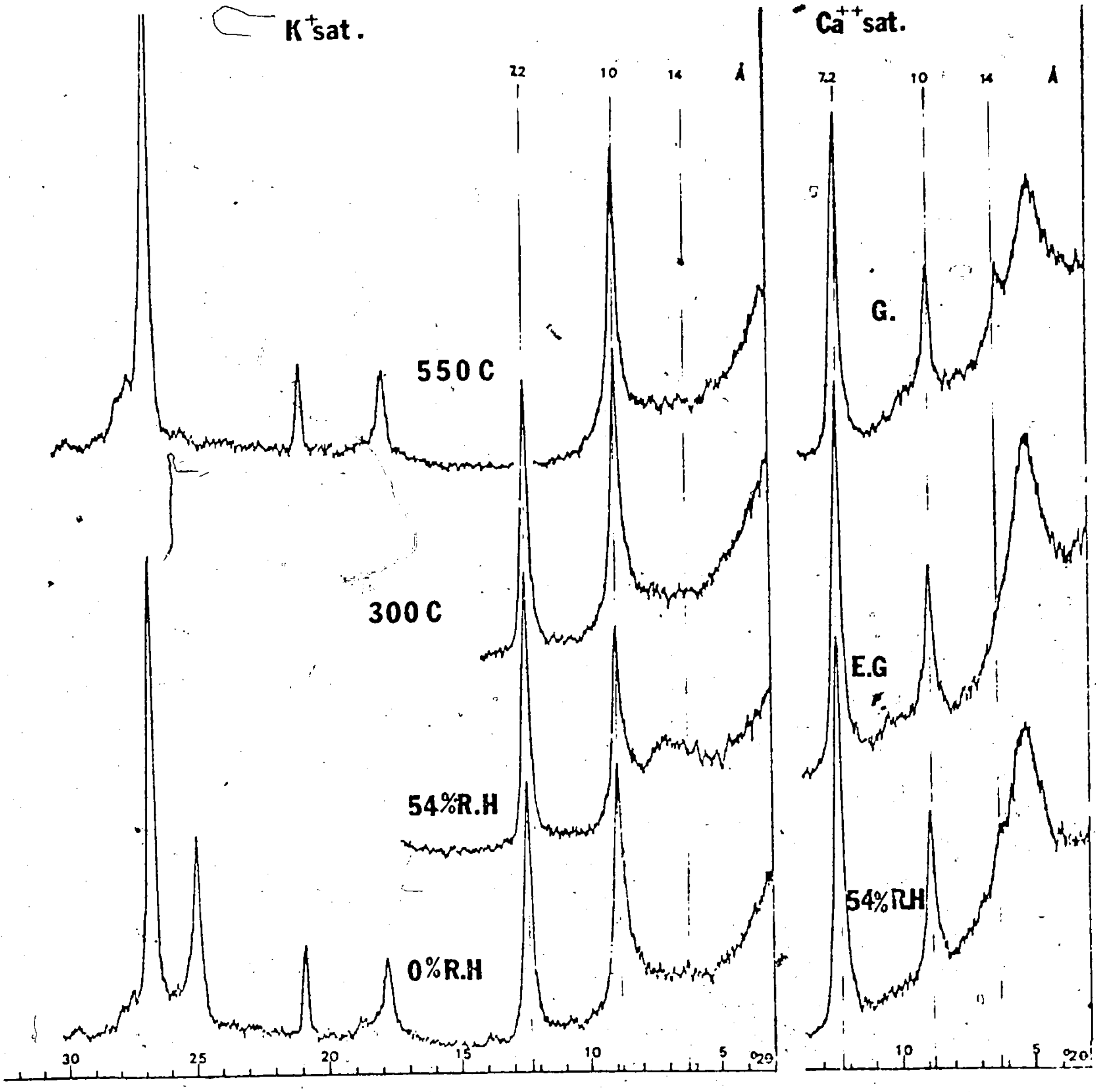


Figure 30. X-ray diffraction patterns of total clay from the Ae horizon of an Orthic Gray Luvisol (site 6)..............................

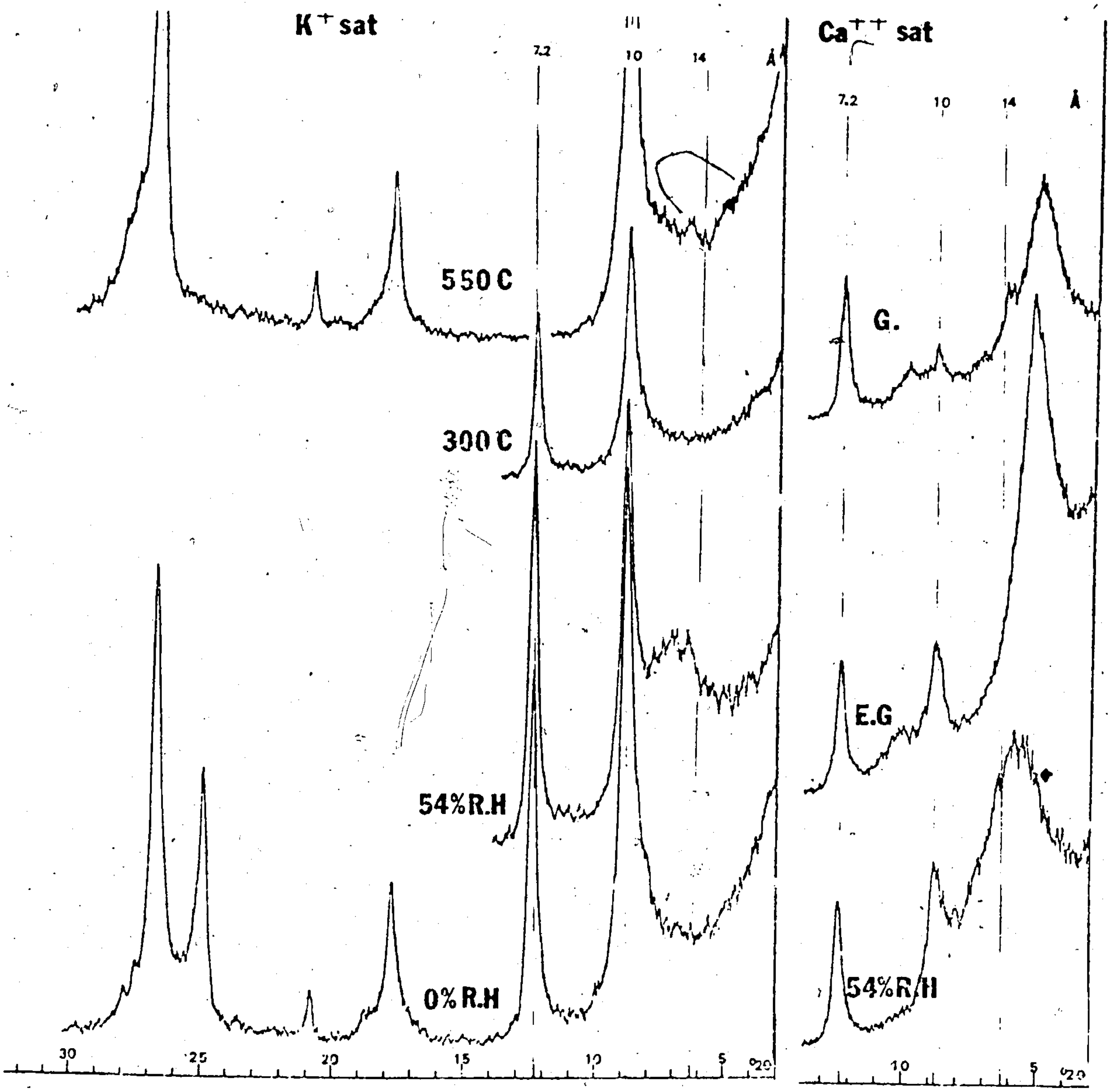


Figure 31. X-ray diffractin patterns of total clay from the Bt horizon of an Orthic Gray Luvisol (site 6)……………………………………

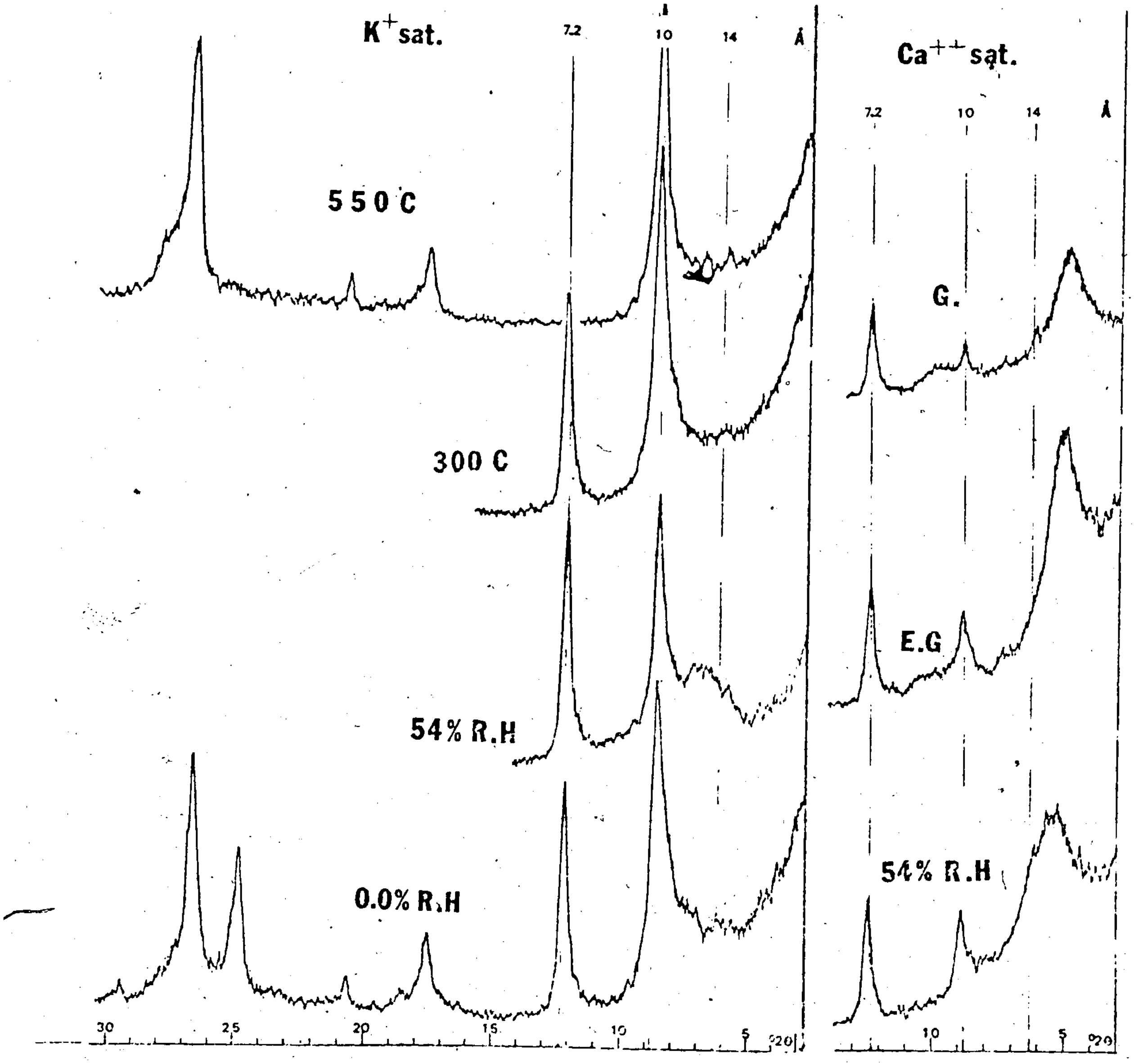


Figure 32. X-ray diffraction patterns of total clay from the Cca horizon of an Orthic Gray Luvisol (site 6)..................... ....

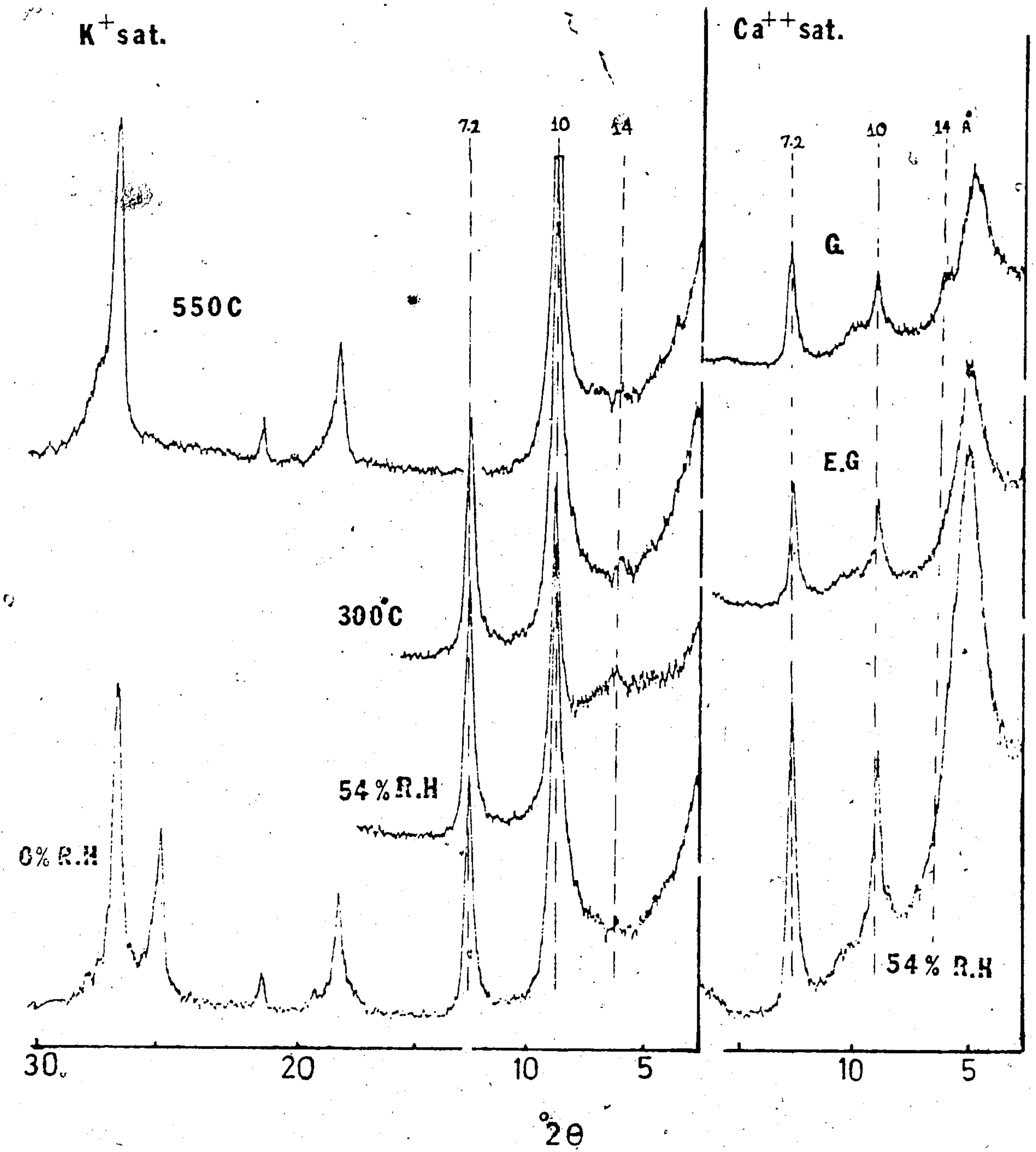


Figure 33. X-ray diffraction patterns of total clay from a shale sample collected from an outcrop in the Sand River area ....