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R.L. Christie
N.J. McMillan

1991

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Scientific editors

R.L. Christie
N.J. McMillan

Editor

L. Reynolds

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Fossil stump (mummified) near Geodetic Hills, Axel Heiberg Island. It is preserved in the Buchanan Lake Formation (Eocene). The stump is about one metre in diameter at the top.

Photograph by Jane Francis

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PREFACE

Fossil forests are ancient forests whose tree stumps have been preserved in growth position. The Tertiary fossil forests of the Canadian Arctic were first described in 1886, but their significance and diversity were not realized until 1955 when Geological Survey of Canada geologists participating in Operation Franklin re-examined them. In 1985, GSC geologists identified another site east of Geodetic Hills on Axel Heiberg Island. Here, as elsewhere in the Arctic, the wood and forest mat are mummified. In some places the wood is so well preserved, it still retains fairly high amounts of original organic matter and looks like modern wood.

This bulletin contains a multidisciplinary study by several scientists, who have addressed the geological setting, the characteristics of the wood and fossil soils, as well as aspects of paleomagnetism, palynology, and paleobotany. Insight is given into how forests adapted to warm high latitudes and accompanying long periods of darkness. The existence of mature forests so close to the north pole in Tertiary time indicates the degree to which global climate has varied in the past, and is significant in our understanding of global change. Although scientists will no doubt add to our knowledge of fossil forests in the future, this publication is a fine, solid foundation for future research.

Elkanah A. Babcock
Assistant Deputy Minister
Geological Survey of Canada

PRÉFACE

Les forêts fossiles sont d'anciennes forêts dont les souches ont été préservées dans leur position de croissance. Les forêts fossiles du Tertiaire de l'Arctique canadien ont été décrites pour la première fois en 1886, mais leur importance et leur diversité n'ont été étudiées qu'à partir de 1955 alors que la Commission géologique du Canada participant à l'Opération Franklin a réexaminé ce phénomène. En 1985, les géologues de la Commission géologique ont identifié un autre site à l'est des collines Geodetic sur l'île Axel Heiberg. Ici comme ailleurs dans l'Arctique le bois et le couvert forestier sont momifiés. Par endroits, le bois est tellement bien préservé qu'il retient une bonne quantité de matière organique originale et ressemble à du bois actuel.

Ce bulletin est le résultat d'une étude multidisciplinaire où plusieurs scientifiques ont bien défini le cadre géologique, les caractéristiques du bois et des sols fossiles, y compris, les domains du paléomagnétisme, de la palynologie et de la paléobotanique. Des éclaircissements sont apportés sur la façon dont se sont adaptées les forêts à la chaleur des hautes latitudes tout en supportant de longues périodes de noirceur. L'existence de forêts adultes si près du pôle nord au Tertiaire indique le degré de variation du climat au cours des temps géologiques et son importance pour notre compréhension vis-à-vis d'un changement aussi global. Les scientifiques, par leurs travaux, nous apprennent ce que seront les forêts fossiles du future et, de conclure, que cette publication par son exposé bien défini nous apporte un fondement solide pour des recherches ultérieures.

Elkanah A. Babcock
Sous-ministre adjoint
Commission géologique du Canada

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FOREWORD

Curt Teichert

Fossil forests of Tertiary age have been known from scattered localities in Arctic Canada for more than 100 years, but the accounts were generally of a reconnaissance nature only. The present volume is the first effort to present a full, multidisciplinary study of a group of fossil forests from high polar latitudes. The various, multifaceted contributions conjure up the picture of a hilly or slightly mountainous landscape at and just below 80° N (paleo- as well as neo-latitude), covered by dense forests of deciduous and evergreen conifers as well as angiosperms. Such forests indicate a temperate to warm temperate climate comparable to that of Florida today, 6000 km to the south. The rich vegetation and presence of large floodplains indicate abundant rainfall, with mild wet winters and warm moist summers. Such an environment is not present at such latitudes today and this scenario poses many questions to which answers must be found. Some assumptions, albeit commonly held ones, are exposed as questionable.

For example, many paleomagnetists, and perhaps also many geologists, believe that there have always been climatic zones similar to those of the present and that their position was determined by the position of the "paleopoles." But if this were true, then the Eocene poles must have been in a position very different from that suggested by Irving and Wynne in this volume. An alternative interpretation would be to conclude that the earth's atmosphere has suffered relatively rapid changes of temperature, worldwide, in the geological past. The implication would be that paleomagnetism is not nearly as good a guide to paleoclimate as it is believed to be at present, and attempts to reconstruct climatic zones on the basis of "paleopoles" are suspect.

The idea that the earth has gone through periods of global high and low temperatures has been expressed in various ways by many authors over the years. Oswald Heer noted in the middle and late 1800s that warm temperate and subtropical floras had flourished during Tertiary times in areas that are now Greenland and Svalbard. Nevertheless, the concept of contrasting climatic zones persists today. Attempts at explaining rapid climatic changes during Tertiary time have been especially vexing and unsatisfactory. Danish scientists have recently discovered a 1.5-2 million year old (latest Pliocene) forest in Peary Land, northern Greenland, which was probably close to the northern tree limit at that time. Since then the northern tree limit has been displaced nearly 2000 kilometres to the south and the time between the extinction of the Peary Land forests and the onset of Pleistocene glaciation in northern Greenland was clearly insufficient for continental movements or polar shifts to account for the climatic change. It seems that future research should be focused more on the possible causes of such rapid global climatic changes. This volume presents abundant material from which to make a start.

SUMMARY

A new fossil forest site, of Eocene age, was found in 1985 near the Geodetic Hills on Axel Heiberg Island, Canadian Arctic Archipelago. The discovery attracted the attention of paleobotanists and other scientists, especially those concerned with paleoclimates and the adaptation of plants and animals to polar conditions of light, season, and climate. This volume comprises 14 papers, many of which describe and discuss the geology and paleobotany of the site and of the forest plants. Other papers consider the probable paleolatitude of the site, the soils in which the ancient forests grew, the microscopic character of the wood, and the fossil fungal forms. A paper on the pollen flora deposited while the forest existed is a key one, and confirms the age of the forest beds as Middle Eocene.

The history of discovery of fossil logs, and especially of fossil stumps in upright, or growth, position, is reviewed by R.L. Christie and N.J. McMillan in the introduction. "Brainard's Petrified Forest", consisting of petrified logs scattered on the surface on Judge Daly Promontory, Ellesmere Island, was discovered during the American *Lady Franklin Bay* expedition of 1881-83, led by Adolphus Greely. Many other wood-bearing localities, some with stumps buried in upright position, were discovered over the years. All of these fossiliferous beds, together with the fossil forest beds near the Geodetic Hills, are included in the Eureka Sound Group, an Upper Cretaceous–lower Tertiary clastic unit of mainly terrigenous sandstone, siltstone, shale, and coal.

B.D. Ricketts describes the geological setting of the forest site. The forest-bearing beds lie within the Buchanan Lake Formation of the Upper Cretaceous – Tertiary Eureka Sound Group. Coarse and fine sediments were shed from a nearby tectonic highland into intermontane troughs and basins: the forests evidently grew on floodplains traversed by sinuous rivers. Sediment accumulation was influenced by the formation of crevasse splays and the migration of channels. Flood events undoubtedly led to the sudden overwhelming and burial of entire forest areas.

In a paper on the dynamics of the fossil forest, J.E. Francis describes the fossil forest with a view to determining the former forest's density of growth and the probable "style" or form of the ancient trees. From measuring the thickness of tree rings, and other data, Francis estimates the productivity of the Eocene fossil forest to have been comparable to that of a mild temperate deciduous forest of today.

J.F. Basinger describes the paleobotany of the Geodetic Hills site, where the in situ stumps and associated forest floor material are extraordinarily well preserved. These forests consisted primarily of deciduous representatives of the family Taxodiaceae, the occurrence of which at a paleolatitude well above the Arctic Circle is taken as evidence of a mild climate. The macroflora reveal a rich regional flora of deciduous conifers and angiosperms, and locally abundant evergreen conifers. *Metasequoia, Glyptostrobus, Pinus, Picea, Larix, Ginkgo, Alnus, Betula, Quercus, Cercidiphyllum, Platanus, Carya,* and others are represented in the fossil remains.

B.A. Lepage and J.F. Basinger review the spatial and temporal distribution of *Larix*, fossil forms of which were found at the Geodetic Hills site. These vegetative remains, which represent the earliest known fossils of this taxon, are known from only a few mummified leaf-litter mats at the site. The distribution of *Larix* in the late Paleogene and early Neogene parallels the spread of coniferous vegetation in the high northern latitudes. The appearance of *Larix* in Europe, however, was evidently delayed until Pliocene time. From a review of the paleogeography of the northern hemisphere, it is concluded that *Larix* became established in Europe through dispersal from Asia when land connections between Europe and Asia were re-established near the end of the Paleogene.

The rich and diverse pollen flora from the sediments enclosing the fossil forests was studied by D.J. McIntyre. He found that forest conditions did not always dominate at the site: mixed woodland and shrubland was an important element of the vegetation for most of the time represented by the forest succession. The pollen assemblages indicate a temperate to warm temperate climate, with little suggestion of subtropical or tropical influence. The beds are assigned a Middle Eocene age, with a Late Eocene age considered possible.

Fossil fungi preserved in leaf-litter layers of the fossil forest succession were examined microscopically by R.G. Day. The fungal spores encountered are mainly members of the largest group of fungi, the Ascomycotina, or their asexual conidial states, the fungi imperfecti (Deuteromycotina). The fossil spores were thought to have a terrestrial character; one complex of propagules was thought to be characteristic of forms typically found in (though not restricted to) temperate climates. In this reconnaissance work, the fungal forms were identified at the generic level only, and little information toward an understanding of the paleoecology of the ancient forest was possible. The fungal spores of the study are illustrated in photomicrographic plates.

The microscopic and chemical characteristics of the buried woods are described in a paper by J.R. Obst and a team at the U.S. Department of Agriculture's Forest Products Laboratory, Madison, Wisconsin (with others). From microscopic features it is clear that the woods have undergone much cellular deterioration. Nevertheless, certain woods could be identified as either spruce (*Picea*) or larch (*Larix*), whereas others may be those of a wide variety of other conifers. One specimen has features indicative of Douglas fir (*Pseudotsuga*). Chemical and instrumental analyses (determination of ash content, wood extractives, acetyl, nitrogen, methoxyl, and alkaline nitrobenzine content, and spectroscopic and spectrometric analyses) were carried out on samples. It is evident that the wood has undergone extensive carbohydrate degradation, with nearly complete removal of hemicelluloses. The residual carbohydrate is mainly crystalline cellulose. The fossil woods are rich in lignins, which are methoxyl-group deficient, side-chain degraded, and more condensed than lignins of recent softwoods. Surprisingly, some wood extractives (e.g., sugars) escaped being released. The Arctic wood was probably degraded mainly through hydrolysis rather than bacterial or fungal decay, evidence of which was not observed.

The petrography of coals in the fossil forest succession is described by F. Goodarzi, T. Gentzis, and B.D. Ricketts. The coals are of low rank (lignite) and consist mainly of the maceral huminite (with humotelinite dominant); liptinite and inertinite are minor components. From ternary diagrams of the various coal components (including mineral matter), the coals can be compared with other, known coals; it appears probable that the coal developed in situ and a forest environment is confirmed. The fossil forest coal is similar to coal of the same, mid-Eocene age at Hat Creek, British Columbia; both coals are low in boron, a feature characteristic of deposition in a freshwater environment.

G.S. Young studied the microscopic characteristics of samples from two fossil stumps using polarized light, phase contrast, and fluorescence microscopy. Slides were prepared using a microtome. The woods have been compacted in burial by horizontal compression but only slightly by vertical compression. Deterioration has been extensive and appears to have resulted mostly from abiotic agents, with only local indication of microbial attack. Hydrolysis in an acidic burial environment is suggested. Microscopic features of the woods, including distinct growth increments and abundant latewood, are consistent with growth having been seasonal between periods of dormancy. Identifications of Taxodiaceae for one of the trees and Pinaceae for the other are supported by the anatomical evidence. Which of the extant genera of the families are represented, however, could not be determined.

Among other distinctive features of the fossil forest beds are paleosols associated with the organic layers. Soil profiles were studied and samples taken by C. Tarnocai and others. Three papers in the volume deal with paleosols. The first, by C. Tarnocai and C.A.S. Smith, describes the paleosols: Podzolic, Gleysolic and Organic. On the basis of what is known of contemporary soil development, the ancient soils are considered to represent a temperate to warm climate with mild, wet winters and warm, moist summers. Certain white layers in the forest succession, earlier taken to be "bleached" zones, were studied closely by Tarnocai, H. Kodama and C. Fox. Their conclusion is that the white layers consist mainly (65% to 98%) of distinctive quartz grains of a platy form. A biological origin for the quartz is proposed; that is, the material probably developed into its present form from opal-phytolites (plant silica).

Soil samples collected early in the study of the fossil forest were studied by A. Foscolos and N.J. McMillan. The clay mineralogy and elemental analyses are reviewed by these authors, who conclude that the buried soil profile studied is only slightly developed, presumably because the forest-bearing floodplains were short-lived geographic features.

The paleolatitude, or the probable latitude of the fossil forests at the time of growth, is reviewed in a paper by E. Irving and P.J. Wynne. Paleomagnetic data for North America, spanning the probable age of the fossil forests (45 Ma), were selected, and the mean paleopole for the interval (55 to 40 Ma) was found to be displaced some 6.6 degrees from the present geographic pole, in the direction of Siberia. Taking errors into account, it appears that the forests grew between latitudes 74 and 80° north; this range is about 2 degrees farther south than the range of present day latitudes of known Eocene fossil forests in the Canadian Arctic. Thus the forests grew at high latitudes, comparable to the latitudes of the sites today. The authors add, however, that the mean ocean temperature at 70° N during the Eocene, determined isotopically, was 10° C; a mean temperature much higher than that of today (-10° C) is thus indicated. Undoubtedly, such warm conditions would have been an important element in forest growth.

The spectacular preservation of the organic material from the fossil forest site is of interest to scientists and nonscientists alike. Because of this, the National Museum of Natural Sciences and the Canadian Conservation Institute (CCI), are investigating the possibilities of preserving the materials for display to museum visitors. D.W. Grattan has conducted various tests, the results of which are given in his paper. He has found that the ancient wood has some of the characteristics of modern wood (e.g., gain and loss of moisture, shrinkage and swelling), but that other characteristics are quite different (e.g., anisotropy in shrinking). The fossil wood in the ground (before exposure to sun and wind) may be essentially waterlogged, with a moisture content approaching 72 per cent. The density is high due to compression (crushing of the cells), and the tensile strength is severely reduced, probably because of the shortness of the cellulose polymer chains. The relative proportions of lignin and cellulose in the ancient wood are comparable to those of old, but historical, waterlogged wood, and the similarity suggests a similar history: immersion in anaerobic bog water for perhaps thousands of years.

Conservation procedures for both the wood and the associated and much more fragile cones, needles and leaves have been developed. For the wood: impregnation with diethylene glycol; for the fragile material: a newly developed sublimation technique of micron-thick polymer film coatings.

SOMMAIRE

En 1985, a été découvert près des collines Geodetic, dans l'île Axel Heiberg de l'Archipel arctique canadien, un nouveau site contenant une forêt fossile d'âge éocène. Cette découverte a attiré l'attention des paléobotanistes et d'autres chercheurs, en particulier de ceux qui étudient les paléoclimats et l'adaptation des végétaux et des animaux aux conditions d'illumination, saisons et climats polaires. Ce volume comprend 14 articles; dans un grand nombre d'entre eux, sont décrites et étudiées la géologie et la paléobotanique du site, et les végétaux de la forêt fossile. Dans d'autres articles sont considérés la paléolatitude probable du site, les sols sur lesquels se sont développés les anciennes forêts, les caractères du bois observés au microscope, et les types de champignons conservés à l'état fossile. Le plus important a été un article relatif à la flore pollinique qui s'est déposée à l'époque de la forêt, article qui confirme que les couches de la forêt fossile datent de l'Éocène moyen.

R.L. Christie et N.J. McMillan font l'historique de la découverte des troncs fossilisés, en particulier des souches qui subsistent en position dressée, ou de croissance. La "Forêt pétrifiée de Brainard" qui se compose de troncs pétrifiés éparpillés à la surface du promontoire Judge Daly dans l'île d'Ellesmere, a été découverte pendant l'expédition américaine de 1881-1883 conduite par Adolphus Greely à bord du *Lady Franklin Bay*. De nombreuses autres localités contenant du bois fossile et parfois des souches ensevelies en position redressée, ont été découvertes au cours des années. Tous ces lits fossilifères, de même que les lits de la forêt fossile proches des collines Geodetic, se situent dans le Groupe d'Eureka Sound, unité clastique du Crétacé supérieur-Tertiaire inférieur, qui se compose principalement de grès terrigènes, de siltstone, de shale et de charbon.

B.D. Ricketts décrit le contexte géologique du site de la forêt fossile. Les lits contenant la forêt se trouvent dans la Formation de Buchanan Lake du Groupe d'Eureka Sound, daté du Crétacé supérieur-Tertiaire. À partir de hautes terres tectoniques proches, se sont déversés dans des cuvettes et bassins intermontagneux des sédiments grossiers et des sédiments fins; de toute évidence, les forêts se sont développées sur des plaines inondables traversées par des cours d'eau sinueux. L'accumulation des sédiments a été influencée par la formation de dépôts de crevasses et par la migration des chenaux. Les épisodes d'inondations ont sans aucun doute provoqué la mort et l'enfouissement soudains de secteurs entiers de la forêt.

Dans un article sur la dynamique de la forêt fossile, J.E. Francis décrit cette forêt, en cherchant à déterminer quelle était la densitéde la forêt, et quels étaient le "style" ou la forme probables des arbres de cette époque. Après avoir mesuré l'épaisseur des anneaux de croissance des arbres et analysé d'autres données, Francis estime que la productivité de la forêt fossile de l'Éocène était comparable à celle d'une forêt actuelle d'arbres à feuilles caduques, formée dans un climat tempéré doux.

J.F. Basinger décrit la paléobotanique du site des collines Geodetic, où les souches en place et les matériaux associés du sol forestier sont extraordinairement bien conservés. Ces forêts se composaient principalement de représentants à feuilles caduques de la famille des Taxodiaceae, dont la présence à une paléolatitude bien plus élevée que le Cercle arctique indique en toute probabilité un climat doux. La macroflore révèle l'existence d'une riche flore régionale de conifères et d'angiospermes à feuilles caduques, et de conifères localement abondants, à feuilles persistantes. Les restes fossilisés sont ceux de *Metasequoia, Glyptostrobus, Pinus, Picea, Larix, Ginkgo, Alnus, Betula, Quercus, Cercidiphyllum, Platanus, Carya* et autres espèces.

B.A. Lepage et Basinger ont examiné la distribution spatiale et temporelle de *Larix*, dont des formes fossiles ont été découvertes sur le site des collines Geodetic. Ces restes de végétaux, qui représentent les fossiles les plus anciens connus de ce taxon, n'ont été découverts que dans quelques tapis de litière de feuilles momifiées, sur le même site. La distribution de *Larix* au Paléogène tardif et au Néogène précoce est parallèle à la propagation des conifères aux latitudes septentrionales élevées. Toutefois, en Europe, l'apparition de *Larix* a de toute évidence été retardée jusqu'au Pliocène. D'une étude de la paléogéographie de l'hémisphère Nord, on conclut que *Larix* s'est établi en Europe par dispersion à partir de l'Asie, à une époque où les liens terrestres entre l'Europe et l'Asie sont réapparus, vers la fin du Paléogène.

D.J. McIntyre a étudié la flore pollinique riche et diversifiée recueillie dans les sédiments contenant les forêts fossiles. Il a constaté que le site n'était pas toujours dominé par des forêts: pendant la majeure partie du temps que couvre la succession de la forêt fossile, des bois mixtes et des forêts d'arbrisseaux représentaient un élément important de la végétation. Les assemblages polliniques indiquent un climat tempéré à chaud, avec peu d'indications d'une influence subtropicale ou tropicale. On estime que ces couches datent de l'Éocène moyen, et peut-être aussi de l'Éocène tardif.

R.G. Day a examiné au microscope des champignons fossilisés, conservés dans des couches de litière de feuilles appartenant à la succession de la forêt fossile. Les spores fongiques trouvées appartiennent principalement à des membres du plus vaste groupe de champignons, les Ascomycotina, ou à leurs formes conidiennes asexuées, les fungi imperfecti (Deuteromycotina). Les spores fossiles ont sans doute un caractère terrestre; un complexe de propagules a paru caractéristique de formes typiquement rencontrées dans des climats tempérés (mais toutefois non limitées à ces climats). Dans ce travail préliminaire d'exploration, les formes fongiques n'ont été identifiées qu'au niveau générique, et il n'existe que peu d'information qui permette de mieux comprendre la paléoécologie de la forêt ancienne. Les spores fongiques examinées dans cette étude sont illustrées dans les planches photomicrographiques.

Un article publié par J.R. Obst et par une équipe du U.S. Department of Agriculture's Forest Products Laboratory à Madison au Wisconsin (et par d'autres chercheurs) décrit les caractéristiques microscopiques et chimiques des bois ensevelis. D'après les détails microscopiques, il est clair que ces bois ont subi une importante détérioration cellulaire. Néanmoins, on a pu identifier certains types de bois comme appartenant soit à des épinettes (*Picea*) ou à des mélèzes (*Larix*), tandis que d'autres pourraient appartenir à de nombreux autres types de conifères. Un spécimen présente des caractéristiques du sapin de Douglas (*Pseudotsuga*). Sur des échantillons, ont été effectuées des analyses chimiques et analyses instrumentales (détermination de la teneur en cendres, produits de distillation du bois, dosage de l'acétyle, de l'azote, du méthoxyle, et du nitrobenzène alcalin, et analyses spectroscopiques et spectrométriques). Il est évident que le bois a subi une importante décomposition des glucides, et une élimination presque complète des hémicelluloses. Le glucide résiduel est principalement de la cellulose cristalline. Les bois fossiles sont riches en lignines caractérisées par la raréfaction des groupes méthanyle, la dégradation des chaînes latérales, et un plus fort degré de condensation que les lignines des feuillus actuels. De façon surprenante, quelques produits extraits du bois (par exemple, les sucres) n'ont pas été libérés. Les bois arctiques se sont probablement décomposés surtout par hydrolyse et non par décomposition bactérienne ou décomposition fongique, ces deux dernières n'ayant pas été observées.

La pétrographie des charbons de la succession de la forêt fossile est décrite par F. Goodarzi, T. Gentzis et B.D. Ricketts. Ces charbons sont de bas rang (lignite) et se composent principalement d'huminite macérale (avec humotélinite dominante); les composants mineurs sont la liptinite et l'inertinite. D'après les diagrammes ternaires des divers composants du charbon (y compris de la matière organique), il est possible de comparer les charbons à d'autres charbons connus; il paraît probable que le charbon s'est formé in situ, et qu'il existait une forêt. Le charbon de la forêt fossile est semblable à des charbons contemporains de l'Éocène moyen, situés à Hat Creek en Colombie-Britannique; les deux charbons sont pauvres en bore, détail caractéristique d'une sédimentation dans un milieu d'eau douce.

G.S. Young a étudié au microscope les caractéristiques d'échantillons prélevés dans deux souches fossiles, en lumière polarisée, en contraste de phase et par fluorescence. On a préparé des lames minces en employant un microtome. Pendant l'enfouissement, le bois a été compacté par compression horizontale, et légèrement seulement par compression verticale. La détérioration a été importante, et semble principalement avoir été causée par des agents abiotiques; il n'existe que localement des indices d'une attaque par des microbes. Une hydrolyse dans un milieu d'enfouissement acide est suggérée. Les détails microscopiques du bois, en particulier des anneaux de croissance bien visibles et du bois final abondant, confirment que la croissance a été saisonnière entre les périodes de dormance. Les indices anatomiques confirment l'appartenance de l'un des arbres aux Taxodiaceae et l'appartenance de l'autre aux Pinaceae. On n'a pu déterminer quels genres actuels de ces familles sont représentés.

Parmi les autres caractères distinctifs des lits de la forêt fossile, citons des paléosols associés aux couches organiques. On a étudié les profils de sols et échantillons prélevés par C. Tarnocai et d'autres chercheurs. Dans le présent volume, trois articles traitent des paléosols. Le premier, rédigé par C. Tarnocai et C.A.S. Smith, décrit les paléosols: podzolique, gleysolique, et organique. De ce que l'on sait sur la formation des sols contemporains, on déduit que les sols anciens représentaient un climat tempéré à chaud, avec des hivers doux et pluvieux et des étés chauds et humides. Certains couches blanches de la succession forestière, considérées autrefois comme des zones "pâlies", ont été étudiées de façon détaillée par Tarnocai, H. Kodama et C. Fox. Ceux-ci concluent que les couches blanches se composent principalement (65% à 98%) de grains de quartz bien définis, de forme tabulaire. On a proposé une origine biologique pour le quartz; c'est-à-dire que le produit existant aurait eu pour origine des phytolites opalisées (silice d'origine végétale).

A. Foscolos et N.J. McMillan ont étudié des échantillons de sol recueillis durant la phase initiale de l'étude de la forêt fossile. Ces auteurs ont étudié la minéralogie des argiles et les analyses élémentaires; ils en ont conclu que le profil du sol enfoui étudié n'est que faiblement développé, sans doute parce que les plaines inondables portant des forêts ont été des détails géographiques de courte durée.

Dans un article qu'ont rédigé E. Irving et P.J. Wynne, ces auteurs étudient la paléolatitude ou la latitude probable des forêts fossiles à l'époque de leur développement. Ils ont sélectionné des données paléomagnétiques relatives à l'Amérique du Nord, couvrant l'âge probable des forêts fossiles (45 Ma); ils ont constaté que le paléopôle moyen de l'intervalle (55 à 40 Ma) était déplacé d'environ 6,6 degrés par rapport au pôle géographique actuel, dans la direction de la Sibérie. Compte tenu des erreurs possibles, il apparaît que les forêts ont poussé entre 74 et 80 degrés N; cet intervalle se situe à environ 2 degrés plus loin au sud que l'intervalle de latitudes où existent actuellement les forêts fossiles connues de l'Éocène, dans l'Archipel arctique canadien. Ainsi, les forêts se sont développées à des latitudes élevées, comparables aux latitudes actuelles des sites considérés. Toutefois, ces auteurs ajoutent qu'à 70° N, durant l'Éocène, la température océanique moyenne déterminée isotopiquement était de 10°C; donc, la température moyenne a été beaucoup plus élevée que la température actuelle (-10°C). Sans aucun doute, de telles conditions climatiques chaudes ont sans doute joué un rôle important dans la croissance de la forêt.

La conservation spectaculaire de la matière organique provenant du site de la forêt fossile est d'un grand intérêt à la fois pour les chercheurs et pour les amateurs. Pour cette raison, le Musée national des sciences naturelles et l'Institut canadien de conservation (ICC) examinent la possibilité de conserver le matériel recueilli pour l'exposer et le présenter aux visiteurs du Musée. D.W. Grattan a réalisé divers tests, dont les résultats sont indiqués dans cet article. Il a constaté que le bois ancien présentait quelques-unes des caractéristiques des bois modernes (par exemple, gain et perte d'humidité, retrait et gonflement), mais que d'autres caractéristiques étaient tout à fait différentes (par exemple l'anisotropie du retrait). Le bois fossile présent dans le sol (avant d'être exposé au soleil et au vent) est parfois gorgé d'eau, et son humidité se rapproche de 72%. Sa masse volumique est élevée, en raison de la compression qu'il a subie (écrasement des cellules), et sa résistance à la traction est fortement réduite, probablement en raison de la faible longueur des chaînes polymères de cellulose. Dans le bois ancien, les proportions relatives de lignine et de cellulose sont comparables à celles de bois gorgé d'eau, ancien, mais datant de l'époque historique, et cette ressemblance suggère une histoire similaire; immersion dans un milieu anaérobie de tourbiére pendant peut-être des milliers d'années.

Des procédés de conservation du bois et des cônes, aiguilles et feuilles associés beaucoup plus fragiles, ont été mis au point. Dans le cas du bois: impégnation avec du diéthylène glycol; dans le cas du matériel fragile: une technique de sublimation nouvellement élaborée, permettant de déposer des enduits polymères dont l'épaisseur est de l'ordre du micron.

INTRODUCTION

R.L. Christie and N.J. McMillan

Christie, R.L. and McMillan, N.J. Introduction. In Tertiary Fossil Forests of the Geodetic Hills, Axel Heiberg Island, Arctic Archipelago, R.L. Christie and N.J. McMillan (eds.); Geological Survey of Canada, Bulletin 403, p. xiii-xvi.

"Fossil forests" are groups of preserved tree stumps found virtually in growth position. They were first discovered in beds of Cenozoic age in the Canadian Arctic a little more than 100 years ago. A new fossil forest site (Fig. 1) was found on Axel Heiberg Island in the Canadian Arctic in 1985 by B.D. Ricketts of the Geological Survey of Canada. The discovery soon attracted the attention of paleobotanists and other scientists concerned with paleoclimates and the adaptation of plants and animals to polar conditions of light, season, and climate. The new fossil forest, which lies near the Geodetic Hills, is now actively being studied by several groups of researchers. The new site is unmatched in terms of completeness of the forest remains and in style and perfection of preservation of the organic matter. From the size and number of stumps, the particular forest species, and the faunules in the enclosing beds, it can be deduced that a temperate, or warmer, climate prevailed at the time the forest thrived. The contrast between the evident conditions of that early Tertiary time and the present conditions at this High Arctic site are striking. Questions naturally arise concerning the effects of polar climate, polar light-and-dark conditions, and the manner in which plants and animals adapt at polar localities (see Francis, 1988).

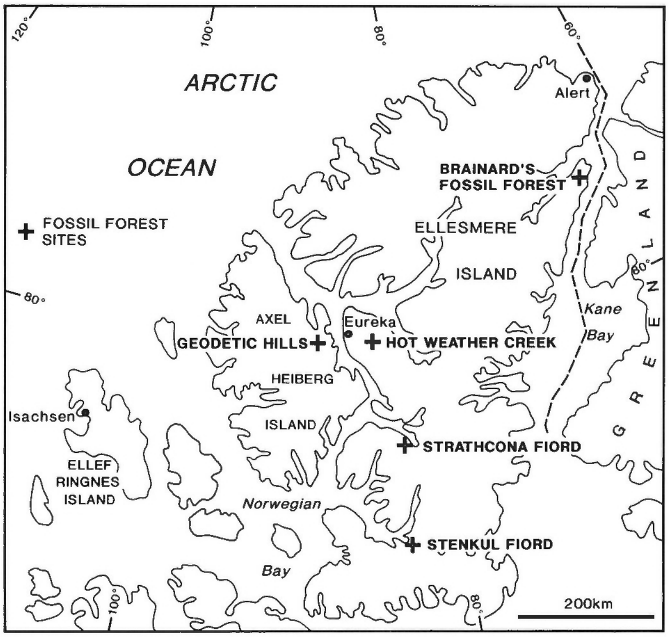

Figure 1. *Fossil forest sites.*

In situ stumps, associated logs and smaller wood fragments, and forest-floor litter are preserved, compressed but without significant petrification or mineralization, at the Geodetic Hills site, northeastern Axel Heiberg Island. Intensive field studies of the forest layers were carried out in 1986 by J.F. Basinger (1986), N.J. McMillan (1986), J.F. Francis (Francis and McMillan, 1987) and others. Further study was undertaken in 1987 and 1988 (see Christie, 1988), when several other researchers joined and undertook specialized studies. The stratigraphy and sedimentology of the enclosing beds has been an ongoing project of the Geological Survey of Canada under B.D. Ricketts.

Several questions have arisen because of the unique style of preservation of the fossil forest layers at Geodetic Hills: did warm temperate, or warmer, conditions exist at this polar latitude in early Tertiary time; was the locality at such a high, or polar, latitude at the time of growth; if darkness prevailed for many months each year, how did the plants maintain lush growth; what can the paleosols tell us about the climatic conditions; and why did the organic matter not rot, oxidize, or petrify during the approximately 45 million years it has awaited exposure in today's Geodetic Hills. The collected papers in this volume form a preliminary account that attempts to answer some of the questions. Undoubtedly, more questions will arise as earlier ones are answered.

The research on the fossil forest sites has been a cooperative venture involving the Geological Survey of Canada, the Polar Continental Shelf Project (both of the Department of Energy, Mines and Resources, Canada), the National Museum of Natural Sciences, and the Canadian Conservation Institute.

Terrigenous plant remains of early Tertiary age were discovered at Watercourse Valley, northern Ellesmere Island, in 1875 by an exploring party under Captain G.S. Nares, Royal Navy[1]. An extensive collection of fossil leaves from this locality was described by O. Heer (1878). Sergeant D.L. Brainard, a member of the Greely expedition of 1881-83, recorded the occurrence of petrified logs on Judge Daly Promontory, south of Watercourse Valley, and the fossil

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[1] For historical accounts, see Christie and Kerr (1981); Christie and Dawes (in press).

locality became known as "Brainard's Petrified Forest" (Greely, 1886). A second "forest" locality, with stumps in growth position, was found at Stenkul Fiord, southern Ellesmere Island, by Per Schei (1903, 1904), of Otto Sverdrup's expedition in the *Fram*. Schei's locality and work were more fully described by Nathorst (1915) and an account of a recent study of the site by C.L. Riediger was published by Riediger and Bustin (1987).

"Fossil forests", with stumps preserved in growth position, were discovered at Hot Weather Creek near Eureka Weather Station, Ellesmere Island, by N.J. McMillan in 1955, during the Geological Survey of Canada's Operation Franklin (McMillan, 1963). A fourth "fossil forest" locality was discovered in 1973 south of Strathcona Fiord, Ellesmere Island, by M.R. Dawson and others (1975). At this site, orange-weathering stumps occurring in black coal provide one of the more spectacular occurrences (see Francis, 1988).

In 1985, a fifth locality, that near the Geodetic Hills, Axel Heiberg Island (the subject of this volume), was found during stratigraphic studies by B.D. Ricketts, Geological Survey of Canada. In situ stumps and root systems, undisturbed forest-floor litter, paleosols with organic and possible phytolite (biological silica) layers, and evidence of dramatic flood deposition of the enclosing sand beds make this an ideal place to study. One of the outstanding features of the Geodetic Hills and the Hot Weather Creek sites is the style of preservation of the wood, leaf, and cone material: much of the organic matter is preserved, not by permineralization, but by processes that allow the appearance and other properties of the organic material to remain close to those of the original materials.

The enclosing beds at the localities at which fossil forests or concentrations of petrified logs have been found were named the Eureka Sound Group by J.C. Troelsen (1950, p. 78). (The Eureka Sound at that time, however, was a "reconnaissance group" in terms of stratigraphic nomenclature.) The Eureka Sound beds – sandstone and mudstone with coal layers, of Late Cretaceous to Oligocene age – have been mapped over the years at widely scattered localities in the Arctic Islands. The formation has in the past been considered dominantly nonmarine in origin, but marine indicators have been found in certain beds (see West et al., 1975, 1977; West and Dawson, 1979). Scattered fragments of petrified wood are characteristic of many of the Eureka Sound terrigenous beds; the wood is typically silicified or calcified, with coaly partings.

Troelsen's reconnaissance unit was later described more fully (and more formally) as a "formation" by E.T. Tozer (1963), and the unit was recently raised to formal group status by two authors, each with a different formational scheme within the Eureka Sound Group. The authors, A.D. Miall (1986) and B.D. Ricketts (1986), worked essentially simultaneously and independently, but the scheme of Miall (1986) has slight priority in time of publication. This development raises some problems in description and communication.

Plant and wood remains found as fossilized detritus (rather than as preserved forest trees and stumps in original position) have been noted at many of the widespread Eureka Sound localities: in a coal-bearing succession at Lake Hazen (Christie, 1964, 1976; Petryk, 1969; Christie and Rouse, 1976; Miall, 1979); as leafy and coaly layers at Bache Peninsula (Christie, 1967); and as plant remains at Yelverton Bay (Wilson, 1979).

Woody remains are also associated with a younger Tertiary detrital unit, the Beaufort Formation (Tozer, 1956). In the initial descriptions of the units, it was suggested that wood of the Beaufort Formation is distinctive in being relatively "fresh" and unmineralized, in contrast to the usually petrified wood of the Eureka Sound Group. Variation in degree and style of preservation of wood, however, has lead to doubts at some localities. For example, at a locality near Lake Hazen, a possible Miocene cone with branches is not petrified (see Blackadar, 1954); on Meighen Island, wood fragments, in situ stumps, and mosses occur in youngest "Beaufort" gravel and sand, with marine silt underlying (see Thorsteinsson, 1961; Fyles, 1962, 1989; Kuc, 1974; Hills and Matthews, 1974).

Where woody remains are absent, the weakly consolidated sands and gravels of either the older, Eureka Sound Group, or the younger, Beaufort Formation are difficult to distinguish; this has led to uncertainty in assignment at several localities. D.G. Wilson (1976) identified both Eureka Sound and Beaufort beds at Yelverton Bay, with the units separated by an unconformity. The Eureka Sound beds of Lake Hazen are capped by weakly consolidated, coarse gravels that were tentatively correlated with Beaufort beds, but which contain coaly units with a pollen flora similar to that of the Eocene of the Geodetic Hills (see Christie, 1976; Christie and Rouse, 1976; Miall, 1979). Indeed, the sand and gravel beds that enclose the fossil forests of the Geodetic Hills, now tentatively accepted as Eocene in age, were earlier assigned Miocene to (?)early Pliocene ages – equivalent to the Beaufort Formation (see Balkwill and Bustin, 1975; Hills and Bustin, 1976; Bustin, 1982).

An understanding is now emerging of the geography, climate, and ecological character of the mid-Eocene forests on Axel Heiberg Island. The region was one of unconnected sedimentary basins with adjacent, rising hilly or mountainous terrains. Sedimentary regimes included terrestrial floodplains, intermontane troughs and basins, and marine basins, some of which were probably small. The climate evidently was mild – temperate to warm temperate – with abundant moisture. Productivity in the ancient forests, in spite of the high latitude and the attendant low sun angles and prolonged periods of darkness, was comparable to that of a warm to hot, temperate, deciduous forest of today. The regional flora included relatively lush lowland or floodplain forests and, probably more widespread, mixed woodland and shrubland. The lowland forests contained primarily deciduous conifers and angiosperms, with locally abundant evergreen conifers.

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SEDIMENTATION, EUREKAN TECTONISM AND THE FOSSIL FOREST SUCCESSION ON EASTERN AXEL HEIBERG ISLAND, CANADIAN ARCTIC ARCHIPELAGO

B.D. Ricketts[1]

Ricketts, B.D., 1991. Sedimentation, Eurekan tectonism and the fossil forest succession on eastern Axel Heiberg Island, Canadian Arctic Archipelago. In Tertiary Fossil Forests of the Geodetic Hills, Axel Heiberg Island, Arctic Archipelago, R.L. Christie and N.J. McMillan (eds.); Geological Survey of Canada, Bulletin 403, p. 1-27.

Abstract

The Middle Eocene fossil forest on eastern Axel Heiberg Island formed within an intermontane basin during the Eurekan Orogeny. Strata that contain the fossil forests (Buchanan Lake Formation) unconformably overlie rocks as old as Early Triassic (Blaa Mountain Group), and as young as the Middle Eocene Iceberg Bay Formation. Thus the climactic period of Eurekan tectonism seems also to have been confined to the Middle Eocene, although a continuation of deformation, perhaps into the earliest Oligocene, cannot be excluded. Therefore, fossil forest strata and associated coarse alluvial deposits record a relatively late stage in the Eurekan tectonic record.

Diabase-rich sediment was shed east and southeast off the rising ancestral Princess Margaret Arch and deposited in settings that ranged from proximal alluvial fans, through braidplains and forest-bearing meanderplains. Mixed hardwood and conifer forests developed on broad floodplains within meander belts that were at least 5 km wide, adjacent to channels several tens of metres wide. The forested meanderplains were 10 to 18 km from the eroding mountain front. The forest succession, consisting of at least 50 lignite/paleosol (Podzol and Gleysol) units within a stratigraphic thickness of about 120 m, records frequent lateral migration of active meander belts during intermontane basin fill. Most of these small-scale cycles formed in response to local conditions of sediment supply with respect to developmental stages of the flora in the swamps and forests. The overall Buchanan Lake succession, however, formed in response to regional tectonism.

Résumé

Dans l'est de l'île Axel Heiberg, la forêt fossile de l'Éocène moyen s'est formée à l'intérieur d'un bassin intermontagneux durant l'orogénèse eurékienne. Les strates contenant les forêts fossiles (formation de Buchanan Lake) recouvrent en discordance des roches qui peuvent remonter au Trias précoce (groupe de Blaa Mountain), et peuvent être aussi récentes que la formation d'Iceberg Bay de l'Éocène moyen. Donc, la période climacique de la tectonique eurékienne semble aussi s'être limitée à l'Éocène moyen, mais l'on ne peut exclure que la déformation se soit poursuivie jusqu'à l'Oligocène initial. Donc, les strates de la forêt fossile et les dépôts alluviaux de granulométrie grossière qui lui sont associés témoignent d'une phase relativement tardive de l'activité tectonique eurékienne.

Les sédiments riches en diabase se sont répandus à l'est et au sud-est, au pied de la proto-arche en surrection de Princess Margaret, et se sont déposés dans des milieux commençant par des cônes alluviaux proximaux, se continuant par des plaines à cours d'eau anastomosés, et aboutissant à des plaines boisées avec cours d'eau à méandres. Des forêts mixtes de feuillus et conifères se sont étendues sur de vastes plaines inondables à l'intérieur de ceintures de méandres d'au moins 5 km de large, à proximité de chenaux de plusieurs dizaines de mètres de large. Les plaines boisées à méandres se trouvaient de 10 à 18 km du front montagneux attaqué par l'érosion. La succession forestière, composée d'au moins 50 unités de lignite/paléosol (Podzol et Gleysol) sur une épaisseur stratigraphique d'environ 120 m, témoigne d'une migration latérale fréquente des zones actives de méandres durant le comblement du bassin intermontagneux. La plupart de ces cycles de petite échelle sont apparus en réponse aux conditions locales d'apports sédimentaires, liées aux étapes de développement de la flore des marécages et forêts. Cependant, la succession globale de Buchanan Lake s'est formée en réponse à la tectonique régionale.

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[1] Geological Survey of Canada, 100 West Pender Street, Vancouver, British Columbia V6B 1R8

INTRODUCTION

A succession of exceptionally well preserved forests occurs in Middle Eocene strata in the Geodetic Hills area of eastern Axel Heiberg Island (Fig. 1). This paper concerns the stratigraphic, sedimentological and tectonic conditions that existed during the periods of forest growth.

The forests grew on broad alluvial plains (Bustin, 1982) situated east of a rising mountain belt, during the climactic period of Eurekan folding and faulting. Stratigraphic successions containing the fossil forests (Buchanan Lake Formation) also provide a record of fluctuating sediment supply that was associated with changing conditions of source uplift (thrusting and folding) and erosion. From a broader perspective, the fossil forest beds form part of a stratigraphic continuum, infilling a narrow, synorogenic, intermontane basin. Similar deposits are known from eastern Ellesmere Island, Otto Fiord, Emma Fiord, Lake Hazen and possibly the Yelverton Bay area of northern Ellesmere. The eastern Axel Heiberg example, how-

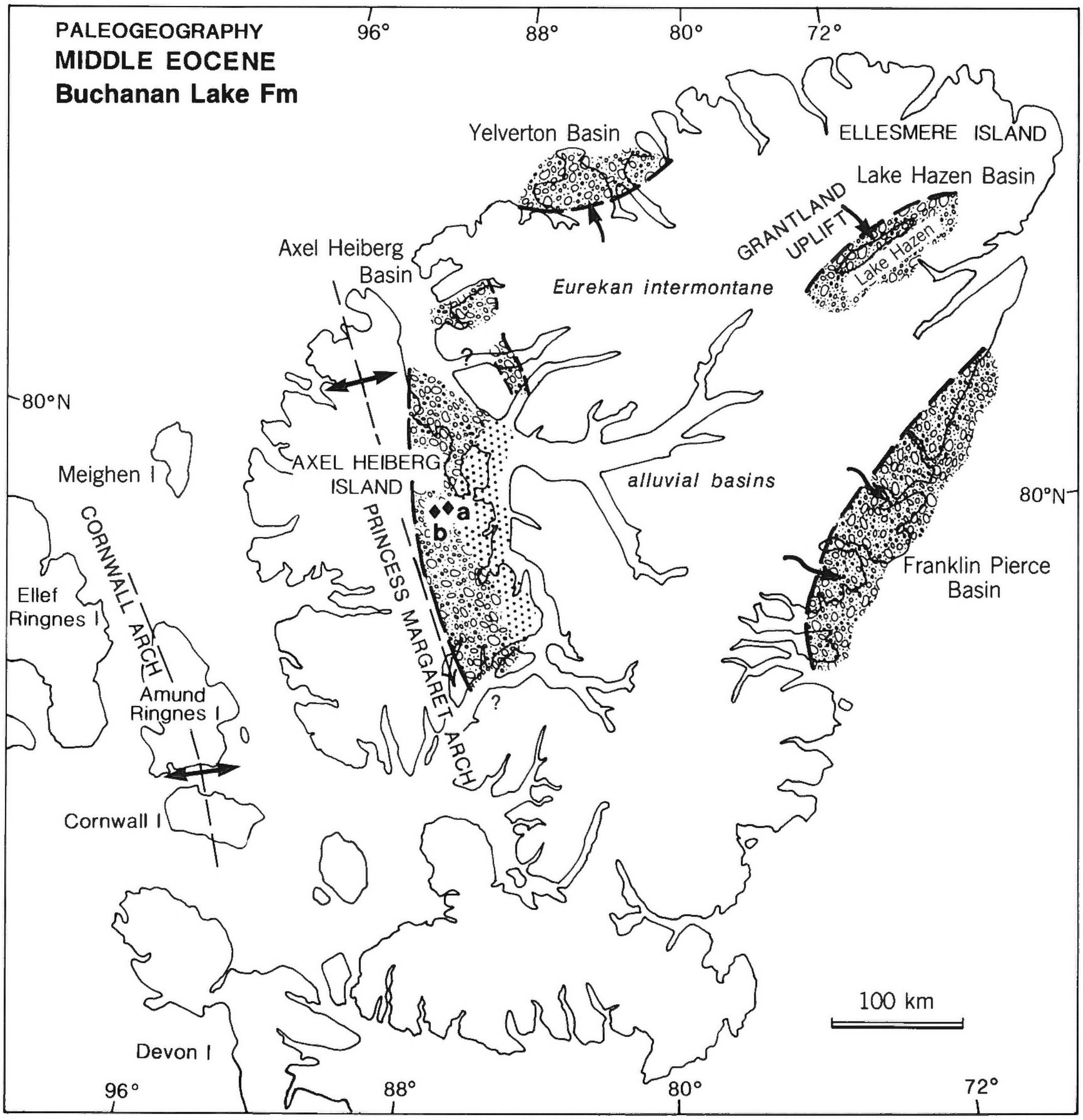


Figure 1. Location of the fossil forest project area, east of Geodetic Hills, eastern Axel Heiberg Island. The forests, within Axel Heiberg intermontane basin, are shown in relation to other Eurekan synorogenic, intermontane basins. In each case the basin limits are approximate. In most cases, the proximal limits of the basins are delimited by large frontal thrusts or uplifts. In general, only the proximal facies are preserved, except in Axel Heiberg Basin, which contains the best representatives of broad floodplain, meanderplain, and forest accumulations. Major Eurekan arches and uplifts are also indicated. The solid diamonds labelled (a) and (b) indicate the approximate centres of aerial photographs shown in Figure 2. Arrows indicate general sediment supply-transport trends.

ever, provides the most complete stratigraphic succession through a Eurekan synorogenic basin, with sedimentary facies that demonstrate clear proximal and distal relationships to the mountainous source terrane.

Prior deliberations

It is quite possible that early Inuit dwellers – and the one or two ancient stone rings or camp sites observed in the area confirm their presence – were the first to discover and perhaps to ponder the fossil forests. Centuries later, mapping during Operation Franklin (1955) placed the forest-bearing strata in a different context of discovery – a geological context. The broad tract of gently sloping land on east-central Axel Heiberg Island was later mapped as Tertiary Eureka Sound Formation by Thorsteinsson (1971). More detailed examination of the stratigraphy and sedimentology in the area east of the Geodetic Hills was undertaken by Hills and Bustin (1976) (see also Bustin, 1982). Strata here were assigned to the Neogene Beaufort Formation, primarily on the basis of a study of spruce cone morphology; cones were compared to Beaufort examples from Banks Island. Subsequent palynological analysis by McIntyre has shown that the low rank coal beds containing the fossil forests are probably Paleogene in age; the bulk of the flora indicates Middle Eocene, but an upper age limit as young as earliest Oligocene cannot be ruled out (see McIntyre, *this volume*). The strata are now regarded as belonging to the Eureka Sound Group, and have been designated the Buchanan Lake Formation (Ricketts, 1986a; Ricketts and McIntyre, 1986).

The depositional setting was interpreted by Bustin (1982) as a system of alluvial fans and adjacent alluvial plains, with detritus eroded from the ancestral Princess Margaret Arch during a period of Neogene block faulting. Although the scenario of timing and style of deformation associated with sedimentation has been modified (Ricketts, 1987a), the overall picture of sedimentation presented by Bustin is valid.

Geomorphology

Broad physiographic subdivisions on Axel Heiberg Island were identified by E.F. Roots (*in* Fortier et al., 1963). The fossil forests of the Geodetic Hills are located near the western limit of the Eureka Sound Uplands, close to the transition to a region described as the Dissected Plateau (Fig. 2). In the study area, the uplands consist of an incised plateau dipping gently northeast, with a maximum elevation of about 600 m immediately east of the Stolz Thrust. This surface appears to be a continuation of the Dissected Plateau (Roots, *in* Fortier et al., 1963) in the central part of the island. The age of the uplands plateau is uncertain; clearly it postdates the Middle Eocene (and possibly younger) Buchanan Lake Formation, but may predate accumulation of the central Axel Heiberg ice cap. A Beaufort age (Neogene) is possible, or alternatively the plateau may have originated during one of the early Pleistocene glaciations.

Drainage is consequent, with a remarkably parallel, northeast-trending pattern. Most of the plateau incision (upward of 300 m) has likely taken place since the major Pleistocene glaciations, in response to isostatic rebound. Flights of river terraces, both erosional (cut into Buchanan Lake strata) and constructional, occur between the forest site and Flat Sound on the east coast.

Strata that contain the fossil forests are best exposed on a pair of prominent, northeast-trending ridges (Fig. 2). Small remnants of the plateau surface occur at about 460 m elevation on one of the ridge tops. The land surface immediately north, south and east of the ridges lies at much lower elevations (about 150 m). Preservation of these ridged uplands may be a result of subtle structural controls; the axes of the ridges approximately coincide with the northeast-plunging axis of a broad, anticlinal warp. Dissection by the consequent drainage has been confined mostly to the flanks of the anticlinal warp. Large braided rivers flank the ridges and are fed primarily by meltwater from an ice cap on Princess Margaret Range. West of the uplands, the braided rivers are transverse to major structural elements such as the Stolz Thrust.

STRATIGRAPHIC AND STRUCTURAL FRAMEWORK

The fossil forest beds form part of the Middle Eocene Buchanan Lake Formation, the youngest stratigraphic unit in the Eureka Sound Group (Fig. 3). Regionally, Eureka Sound Group rocks overlie the Mississippian to Upper Cretaceous Sverdrup Basin and lower Paleozoic strata of the Franklinian Mobile Belt. Deposition of quartz sand, beginning in the middle Campanian, signalled the final stages of thermally driven subsidence that had characterized the underlying and long-lived Sverdrup Basin (corresponding to the earliest phase of Eurekan tectonism proposed by Balkwill, 1978). In the eastern Arctic, Eureka Sound Group rocks, representing a variety of sedimentary environments such as wave- and fluvial-dominated deltas, barrier islands, estuaries, lagoons, coastal plains and shelf deposits, accumulated in a basin known as Fosheim Basin (formerly called Fosheim Foredeep by Ricketts, 1987a). During the Late Cretaceous, the locus of sedimentation in Fosheim Basin coincided approximately with the earlier depocentre in Sverdrup Basin, but the basin expanded during the Paleogene to encroach upon Franklinian bedrock.

During the Middle Eocene, the entire eastern Arctic region underwent a fundamental structural reorganization. Crustal shortening and deformation accompanied the opening of Baffin Bay and Labrador Sea by seafloor spreading, and the counterclockwise rotation of Greenland against the northeastern margin of the North American craton (Srivastava, 1985).

Major west to northwest directed compression gave rise to the climactic phase of Eurekan tectonism, producing foreland-style folding and faulting. Analysis of the relative motion between the North American and Greenland plates indicates that the main phase of deformation began at some time between magnetic anomalies 24 through 21, corresponding to

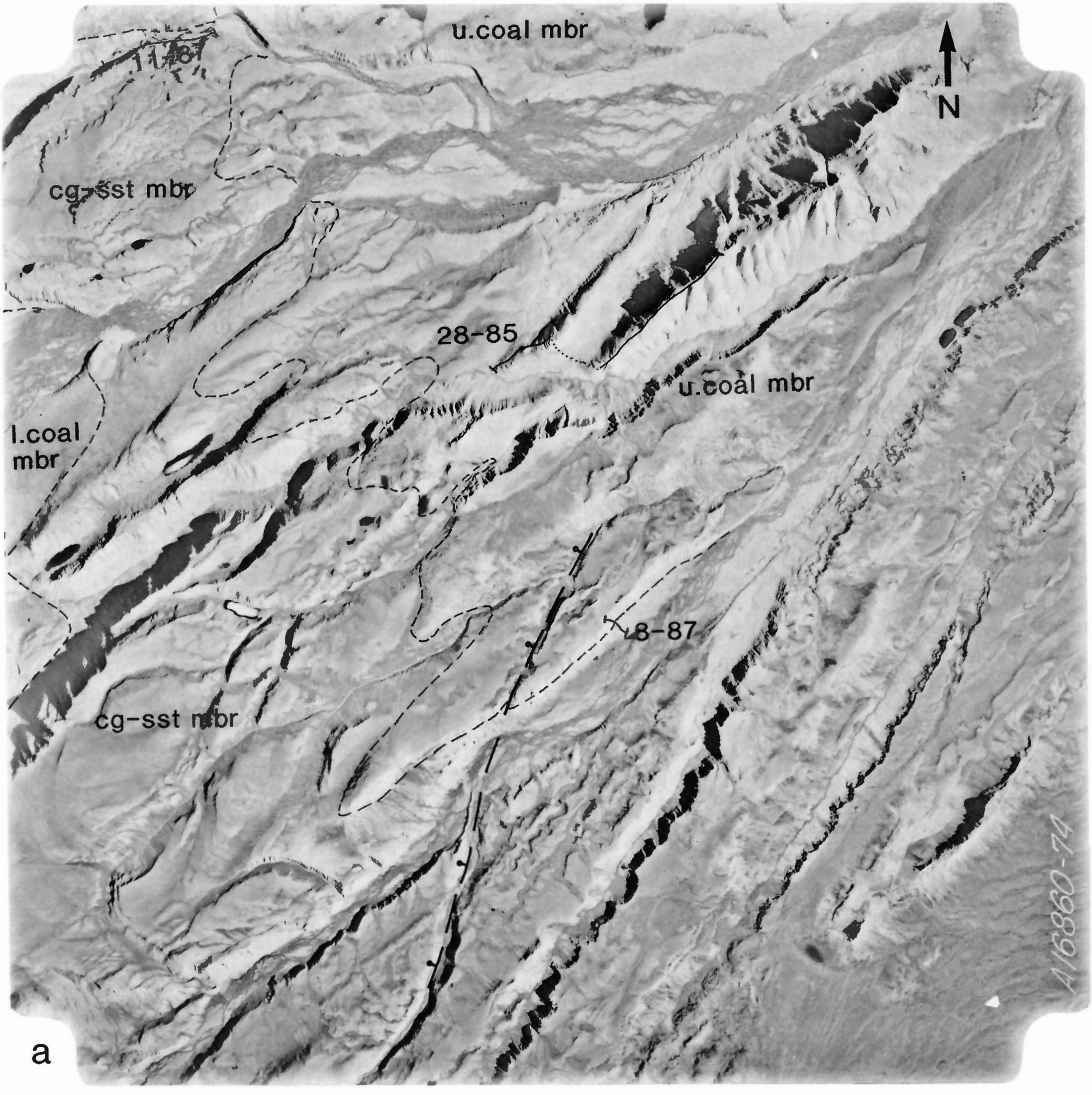
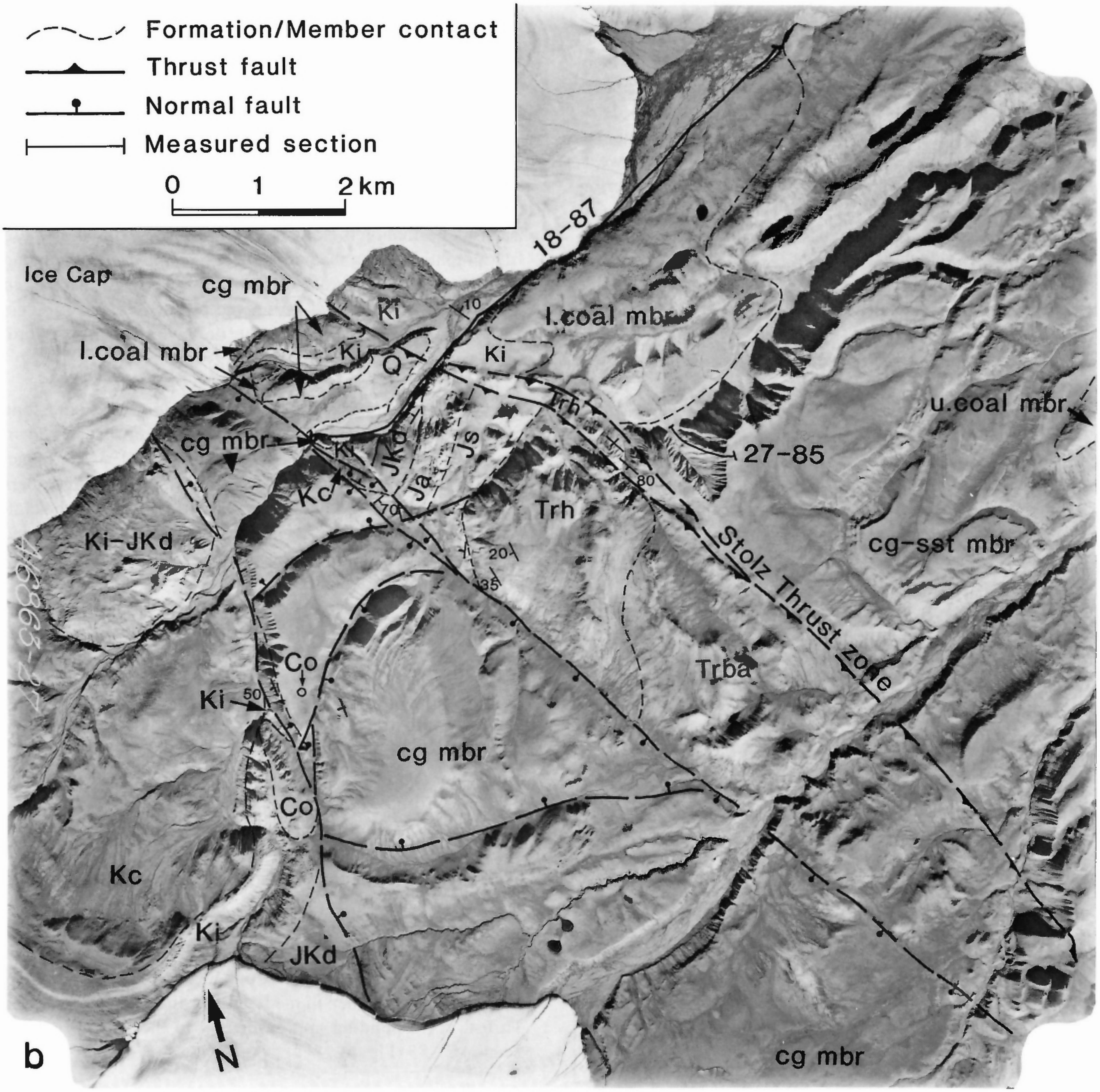


Figure 2. Vertical (N.A.P.L.) aerial photographs of A, the fossil forest locality, and B, west to the hanging wall of the Stolz Fault Zone, with measured sections and map units indicated. Modified from Thorsteinsson (1971), Bustin (1982), and Ricketts (1987). The photo centre of Fig. 2a (A16860-74) is approximately 79°54'N; 89°00'W, whereas Fig. 2b (A16863-27) is approximately 79°51.5'N; 89°25'W. Map units of the Buchanan Lake Formation are designated: cg mbr = conglomerate member; l. coal mbr = lower coal member; cg-sst mbr = conglomerate-sandstone member; u. coal mbr = upper coal member. Q = Quaternary. Older Sverdrup Basin units are: Kc = Christopher Fm; Ki = Isachsen Fm; JKd = Deer Bay Fm;

major changes in the kinematics of Arctic seafloor spreading and to significant global plate reorganization (Rowley and Lottes, 1988; Scotese et al., 1988). By anomaly 13 (about 35 Ma – earliest Oligocene) only small amounts of relative plate motion between Ellesmere Island and North Greenland could be determined (Rowley and Lottes, 1988), and hence Eurekan



Ja = Awingak Fm; Js = Savik Fm; Trh = Heiberg Fm; Trba = Blaa Mountain Gp; Co = Diapiric anhydrite and gypsum. With the exception of two small outliers adjacent to (south of) the main ice cap, the conglomerate member in Figure 2b is confined to a graben southeast of the Stolz Thrust. Stratigraphic contact between the conglomerate member and Isachsen Formation is exposed in the structurally complex area between the graben and the trace of the main Stolz Thrust. In the footwall (east) of Stolz Thrust, the basal unit is the lower coal member that overlies Isachsen sandstone and coal. A small anhydrite diapir (Co) has intruded the conglomerate member (Fig. 2b, left centre).

tectonism was likely completed by this time. There is, however, no independent biostratigraphic evidence to support unequivocally the hypothesis that sedimentation continued until the Oligocene.

Folding and thrusting of Sverdrup Basin and Eureka Sound Group rocks (Fosheim Basin) along the backbone of Axel Heiberg Island, and the production of the ancestral Princess Margaret Arch, fragmented these older basins. Clastic detritus, composing the Buchanan Lake Formation, accumulated on the flank of this uplifted terrane in a narrow, northwest-trending, synorogenic intermontane basin that was superimposed upon the older stratigraphic foundations (Fig. 3). At least four syn- to late-orogenic basins of this kind are known in the eastern Arctic Islands (Fig. 1).

BUCHANAN LAKE FORMATION

Four map units are recognized from the fossil forest site west to the hanging wall of Stolz Thrust (Figs. 2, 3). In ascending order, these units are: the *lower coal member*, the *conglomerate member*, the *conglomerate-sandstone member*, and the *upper coal member*. The fossil forest beds that are the primary subject of this volume are contained in the upper coal member. Most of the area encompassed by the aerial photographs in Figure 2 was traversed, and several stratigraphic sections were measured in detail (Fig. 6).

The lower coal member has not been reported previously. Lithologically, it is similar to the upper coal member, consisting of interbedded lithic sandstone, mudstone and lignite, commonly arranged in fining-upward sequences. Tree stumps in growth position and abundant leafy material in the mudstones and lignites generally indicate forested floodbasin environments. The most extensive exposure of the lower coal member is along the right bank of an unnamed river that flows along the margin of a prominent lobe of the ice cap (Fig. 4).

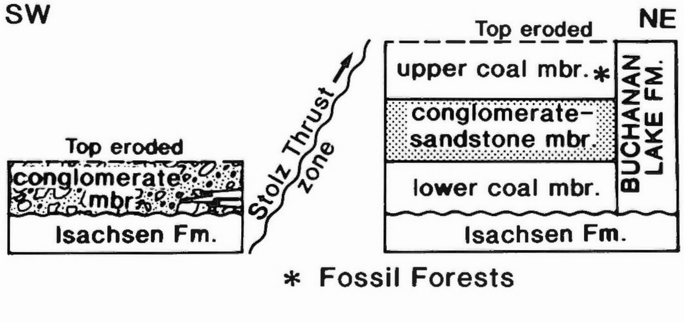


Figure 3. *Schematic summary of lithostratigraphy of the fossil-forest-bearing strata, Geodetic Hills area. Note that the conglomerate member in the hanging wall of the Stolz Thrust contains an interfingering sliver of lower coal member strata.*

Additional outcrops occur west of the main Stolz Fault (i.e., in the hanging wall – Fig. 2b), where they appear to interfinger with the conglomerate member. The maximum thickness of the lower coal member is estimated to be about 350 m (the total thickness of the Buchanan Lake Formation in the footwall of the Stolz Thrust is greater than earlier reported, approaching 850 m). Underlying these strata, and immediately east of the main fault, are 70 to 80 m of white, quartz sandstone and thin coal seams of the Isachsen Formation (Fig. 2b).

The conglomerate member is exposed in a graben west of the Stolz Thrust. This member interfingers at its base with a thin wedge of the lower coal member. Its contact with the Isachsen Formation is exposed in a few places. Based on these stratigraphic criteria, the conglomerate member in the hanging wall is tentatively correlated with the lower coal member in the footwall. In terms of sedimentological relationships, the coal-bearing strata could be the downslope alluvial plain equivalents of the conglomeratic alluvial deposits.

Lithostratigraphic details of the conglomerate-sandstone member and upper coal member have been described in Ricketts (1986a) and Ricketts and McIntyre (1986). Contact between the lower coal and conglomerate-sandstone members is gradational over a few metres (Fig. 5), as is the contact at the base of the upper coal member. Representative stratigraphic sections of the conglomerate-sandstone and upper coal members are shown in Figure 6.

FACIES ASSOCIATIONS AND SEQUENCES

The Buchanan Lake Formation can be described in terms of seven lithofacies associations, some of which exhibit sensible[1], sequential changes in internal organization (Table 1). Most of these facies associations are recognizable in other outcrop areas of the Buchanan Lake Formation, such as Emma Fiord, Otto Fiord, Stang Bay, Mokka Fiord and the eastern coast of Ellesmere Island.

Matrix-supported conglomerate association

This facies association has been identified at two localities in the Geodetic Hills area within the conglomerate member west of the Stolz Thrust, and at one locality in the conglomerate-sandstone member immediately in the footwall of Stolz Thrust, corresponding approximately to the 280 m level in Section 27-85 (Fig. 6).

Very coarse, matrix-supported conglomerate occurs in beds up to 10 m thick at the western limit of exposure of the Buchanan Lake Formation (Figs. 2; 7A,B). They are interbedded with units of the clast-supported thick conglomerate

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[1] Capable of being sensed.

association (Fig. 7B) and are probably the most proximal facies preserved in this part of the basin. Eighty-five to ninety per cent of the clasts are diabase and gabbro, up to 80 cm in diameter, but averaging 25 to 30 cm. The matrix is a muddy, diabase-rich sandstone.

Two beds in the conglomerate-sandstone member, 1.2 m and 0.8 m thick respectively, and separated by a muddy sandstone veneer, are composed of very poorly sorted, matrix-supported monomict conglomerate. The contact between these beds is locally eroded. Angular to subrounded clasts range in size from a few millimetres to 32 cm, and consist almost entirely of white, indurated sandstone, some having bioturbation and relict crossbedding. No diabase pebbles were observed. This facies association directly overlies the conglomerate-sandstone association.

Thick conglomerate association

Tabular units of cobble and boulder conglomerate and thin recessive sandstone beds, 3 to 10 m thick form prominent ribbed exposures in the conglomerate member west of the Stolz Thrust. The thin sandstone interbeds generally make up less than 25 per cent of the units and, although exposures tend to be rubbly, seem to be part of fining-upward packages. Moderately rounded pebbles and cobbles up to 45 cm across (averaging 8 to 10 cm), form a clast-supported framework. Internally, the tabular bedded units are crudely layered, but in places contain large wedge sets. Sorting is poor. Some imbrication occurs, indicating eastward flow, and wood fragments are concentrated in lags. Bustin (1982) also provided sedimentological details of this lithofacies.

Conglomerate-sandstone association

Tabular bedded units of crudely stratified conglomerate, up to 10 m thick, form about 10 to 20 per cent of this association (the remainder consists of 50 to 60 per cent crossbedded conglomerate units and 30 to 40 per cent sandstone beds). Individual beds can be traced along the structural dip for 50 to 80 m. The overall range in clast size is 1 to 50 cm with an average of 10 cm, although within specific bedding units the

Figure 4. *Exposure of the lower coal member adjacent to the ice cap (see Figs. 2 and 5 for location). The cliff section is about 15 m high and contains several fining-upward sandstone-coal and siltstone-coal units.*

size range is less and sorting is moderate. The coarsest component consists of indurated quartz sandstone, whereas clasts of diabase generally make up a slightly finer fraction. Proportions of diabase range from 20 to 80 per cent. Diabase-rich conglomerates tend to occur in the upper stratigraphic levels of the conglomerate-sandstone association, although exceptions occur locally. Clast imbrication indicates flow toward the east (Fig. 8). Thin sandstone lenses occur locally, but these are generally more common and of greater lateral extent at the bounding surfaces of tabular bedded units (Fig. 9).

The sandstone lenses, which are clearly wedge shaped in three-dimensional exposures, range from a few centimetres to 4 m thick and discordantly overlie the conglomerates. They range in outcrop width from less than one metre to 30 m and more. Ripples, and planar-tabular crossbeds in sets up to 50 cm thick, are common. Crossbed sets may be lined with a veneer of pebbles. Sandstone wedges may be stacked, each wedge separated by a pebble veneer or a small wedge of conglomerate (Fig. 10). In places, small channel-like scours occur, some filled with mudstone plugs.

Crossbedded conglomerates in tabular and wedge shaped bedding units a few centimetres to 3 m thick extend up to 50 m laterally, and constitute 40 to 50 per cent of the facies association (Fig. 11). Clast fabrics and composition are similar to the tabular, crudely stratified conglomerates. Planar-tabular crossbed foresets dip 10 to 15 degrees, display normal and updip size grading, and are locally veneered with sandstone. Although the average clast size is similar to the crudely stratified conglomerates, the size range is generally less. Contacts between conglomerate wedge sets are commonly defined by sandstone lenses like those described above (Figs. 9, 11).

Conglomerates that occupy distinct channel structures were encountered in a few sections, usually cut into sandstone. Channel widths range from 5 to 10 m. Larger structures may

Figure 5. *Panorama of the lower coal member (mainly below the ice cap) and the overlying conglomerate-sandstone member (c-s) – the cliff-forming unit in the background (viewed toward the northeast). Photograph taken immediately above cliff exposure in Figure 4 (arrow).*

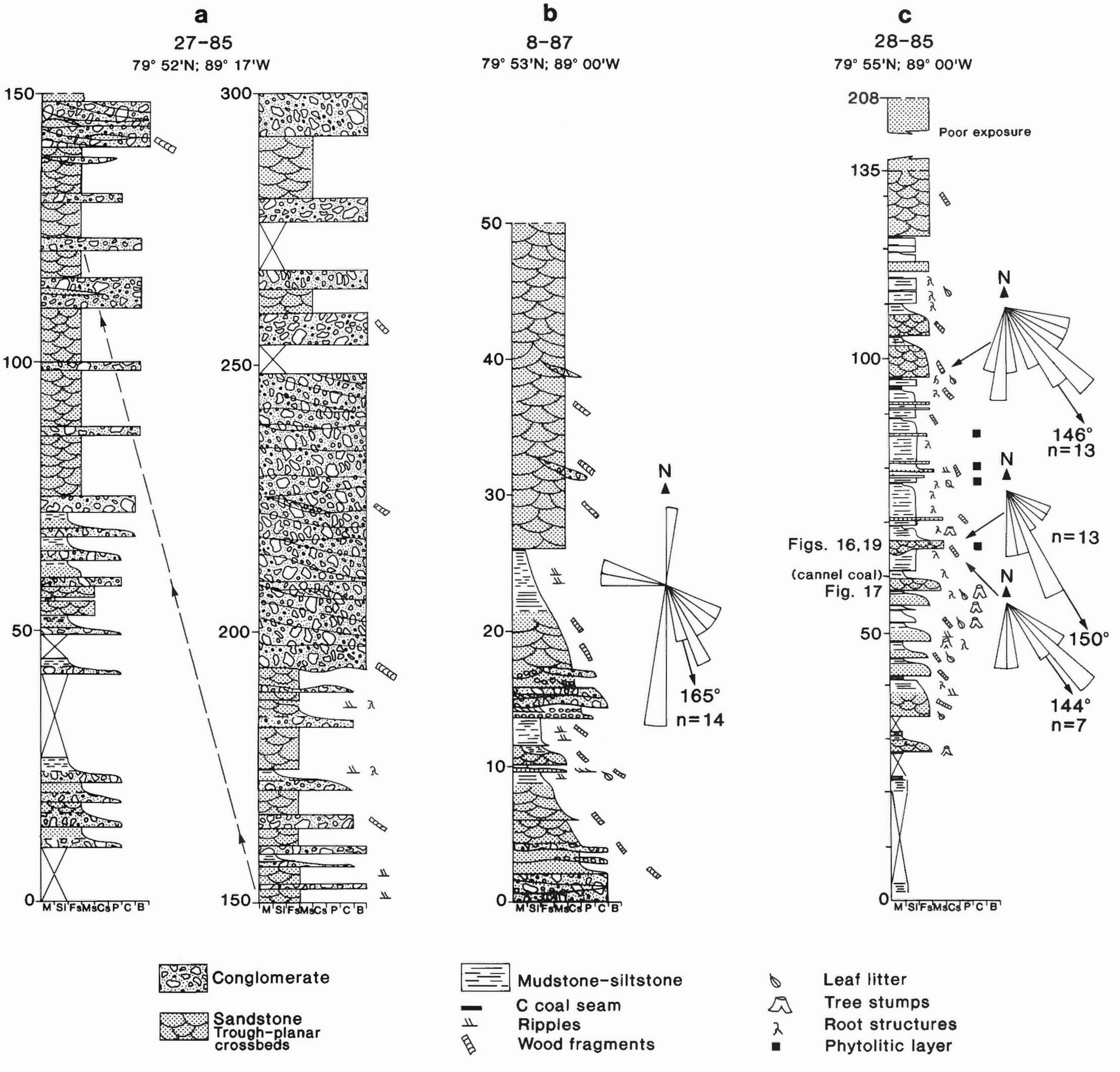

Figure 6. Schematic representation of measured sections of the conglomerate-sandstone and upper coal members. Section locations are shown in Figure 2. Section thicknesses are in metres. The Wentworth grain size scale is used. Paleocurrent rose diagrams in b and c are based primarily on trough crossbeds; for each, the vector mean direction and number of measurements (n) is given. In b, the crossbed azimuths encompass the 50 m of section; in c, the measurements at specific levels are indicated by arrows. In a, the facies is mostly braidplain with possible distalmost alluvial fan. b shows transitional facies of gravelly meandering and possible anastomosing streams. (Note the mud plug at 11 m.) Column c shows a cyclical succession of meandering stream, crevasse splay and floodplain deposits with many fossil forest units. The phytolitic layers seem to be confined to about the 65 to 95 m interval in c.

be present, but the vagaries of exposure do not permit ready identification.

There is no obvious regular facies sequence at the scale of individual beds in the conglomerate-sandstone association. However, over a stratigraphic thickness of 50 to 80 m, there are overall fining-upward trends, where the proportion of sandstone increases to 35 or 40 per cent, accompanied by a concomitant fining of the conglomerate framework and an increase in the preponderance of tabular crossbedded conglomerate. Along one transect (Section 11-87, Fig. 2), this fining-upward trend was repeated three times; however, it is not possible to correlate these large fining-upward sequences in adjacent transects.

Thick sandstone association

Successions of medium grained, moderately sorted, brown lithic sandstone, 20 m and more in thickness, constitute a facies association quite distinct from the sandy component of the conglomerate-sandstone association (Fig. 6). The lateral extent of this association is not known. The internal organization is characterized by abundant planar-tabular and trough crossbedding in sets 20 to 60 cm thick. Some trough bases are lined with mudchip lags and a few pebbles of indurated sandstone or diabase. Plant debris is common.

Crossbedded sandstones are arranged in stacked sequences 1 to 10 m thick, each delineated by a scoured or channelled base. Some scour fills are packed with tangled logs and twigs. There are no indications of fining- or coarsening-upward trends.

Paleosols

Although coal (lignite) seams are usually thin and sparse in this facies association, some Podzols with brown B horizons are developed (Tarnocai and Smith, *this volume*) in association with the lignites or at discordances in the crossbedded sandstone.

TABLE 1

Facies associations of principal bedforms

Facies associations	Principal bedforms	Lithostratigraphy
Matrix-supported conglomerate	Very poor sorting Lacking internal layering Generally nongraded	Conglomerate member; rarely conglomerate-sandstone member
Thick conglomerate	Tabular bedded units Crude stratification Clast imbrication Rare crossbedding	Conglomerate member
Conglomerate-sandstone	Tabular bedded units Horizontal stratification Planar-tabular crossbedding Channel fill Clast imbrication	Conglomerate-sandstone member
Thick sandstone	Stacked planar-tabular and trough-crossbedding Ripples Scour and fill	Conglomerate-sandstone member
Conglomerate-mudstone (coal)	Fining-upward units Basal pebble lag or conglomerate bed Planar-tabular and trough crossbedding	Conglomerate-sandstone member
Sandstone-coal	Fining-upward units Channel fill Lateral accretion foresets (epsilon, or inclined heterolithic bedding) Prominent lignite seams	Lower and upper coal members
Siltstone-coal (paleosol)	Fining-upward units Prominent lignite seams	Lower and upper coal members

Figure 7A. *Thick, poorly stratified, matrix-supported conglomerate facies in the conglomerate member, exposed at the western limit of the Buchanan Lake Formation near Geodetic Hills. The largest boulders here are 80 cm across. Sandy-mud matrix comprises 20 to 25 per cent of the breccia. About 80 per cent of the clasts are diabase or gabbro.*

Figure 7B. *Steeply dipping units of matrix-supported conglomerate (clast-supported) interbedded with thin, discontinuous sandstone beds of probable stream flood origin. Same locality as Figure 7A.*

Conglomerate-mudstone (coal) association

A variety of facies are arranged into distinct fining-upward units up to 12 m in total thickness (Fig. 6). Erosional relief at the base of each unit is upward of 50 cm. The conglomerate component, about 20 per cent of unit thickness, consists of parallel stratified, planar-tabular and trough crossbeds, numerous sandstone drapes and lags of plant debris. Contact with the overlying sandstones (40 to 60 per cent of each fining-upward unit) is gradational. A few planar-tabular crossbeds with graded foresets occur in the lower part of the sandy component. Trough crossbeds predominate in the remaining sandstone, where grain size, the abundance of pebble lags and crossbed set thicknesses decrease upward. Laminated and rippled siltstone and mudstone capped by thin, woody lignites (20 to 30 per cent of unit thickness) complete the fining-upward sequence. Some sequences are truncated and exhibit incomplete fining trends. Locally, scours up to one metre deep are filled with carbonaceous mudstone, a basal pebble lag, and sandy layers containing climbing ripples.

Sandstone-coal association

This, and the closely related siltstone-coal facies association (described below) are best exposed on the ridges containing the fossil forest sites. Although superficially similar to the conglomerate-mudstone association, there are important differences in internal organization. Instead of a basal conglom-

Figure 8. Well developed, east-directed (left) imbrication in crudely stratified conglomerate facies (conglomerate-sandstone association). Scale (ballpoint pen) is 14 cm long. (Section 11-87, Fig. 2a.)

erate, the lowermost sandstones abruptly overlie lignites of the subjacent sequence, with some erosional scouring (Fig. 12). Tree stumps in growth position are buried. A few planar-tabular crossbeds occur near the base, but trough crossbeds are most common, with sets up to 60 cm thick. Crossbed set thickness decreases upward with a concomitant decrease in grain size (Fig. 13). The sandstone grades upward into silty lithotypes, where ripples and parallel laminae predominate, interbedded with thin mudstone layers. The fining-upward sequence is usually capped by lignite seams ranging in thickness from a few centimetres to 1.5 m; thicker seams commonly contain up to 45 per cent interbedded mudstone (Fig. 14). In places, the siltstone-coal component accounts for 40 per cent of the fining-upward unit thickness, almost twice that in the conglomerate-mudstone association. Root structures penetrate to depths of 1.5 m. Tree stumps in both prone and growth positions are also common. It is these lignites that contain the beautifully preserved forest litter flora, characterized by abundant *Metasequoia*, spruce, and hardwoods.

Figure 9. Sandstone veneers and thin lenses mark the contacts between wedges of planar-tabular crossbedded conglomerate facies in the conglomerate-sandstone association. (Section 11-87, Fig. 2a.) Imbrication (indicating eastward flow) is less well developed than in Figure 8. Scale (ballpoint pen) is 14 cm long.

The lateral extent of individual fining-upward units is most easily established by tracing the lignite beds (Fig. 15). To the southeast of the main fossil forest locality fining-upward units can be traced almost 2 km, whereas toward the northeast they can be traced almost 5 km (limited by the extent of outcrops). Changes in seam thickness over these distances amount to only a few per cent. However, changes in the thickness of associated sandstone sequences are as great as 30 to 40 per cent, and in some cases this is delineated by the presence of channel structures. The example illustrated in Figure 16 shows that about half of the channel cross-section is preserved (15 m wide) with a maximum 4 m thickness of sandstone fill. The channel base is discordant on underlying crossbedded sandstone and is lined with comminuted plant debris; in places the channel floors are littered with logs (Fig. 17). Along strike, the same channel truncates mudstone, lignite, and a distinctive white paleosol. Base-of-channel discordances such as these are common in the sandstone units; however, they tend to be laterally extensive rather than vertically stacked. Crossbed directions within the channel fill indicate consistent southeast-directed flow (Fig. 6).

An example of laterally accreted foreset bedding, or epsilon bedding, occurs in the left bank of an unnamed river at Section 11-87 (Figs. 2a, 18). Set thickness is about 4 m, and outcrop extent is about 100 m. The basal contact appears planar (the outcrop is inaccessible) and truncates two thin lignite beds. Foresets consist mostly of sandstone with minor interleaved mudstone drapes, dipping about 15 degrees in the lower part of the accretionary set and decreasing upward; some foresets are sinusoidal. A thin lignite bed caps the unit.

Paleosols

The sandstone-coal facies association is characterized by Podzolic paleosols with well developed B horizons containing

Figure 10. *Details of structure in pebbly sandstone layers between crossbedded conglomerate sets. Stream flow was to the left, established from clast imbrication and other bedforms in this unit. The central, oblate cobble formed a local obstruction to flow; wherein the underlying layer of pebbles gradually accumulated during progressive scouring around the margins of the cobble. The scour face is steepest on the upstream side (right) and shallower on the lee side. Plant debris was trapped on the upstream face of the cobble. Scale on the right is 5 cm long. (Section 11-87, Fig. 2a.)*

concretions; a thin, leached A horizon; and (where present) a thin Organic cap layer that usually is permineralized (Tarnocai and Smith, *this volume*). These red-brown Podzols are up to 150 cm thick.

Siltstone-coal (paleosol) association

These fining-upward units, 0.5 to 4 m thick, consist predominantly of siltstone, mudstone and lignite in approximate proportions of 50, 40 and 10 per cent, respectively. They consistently lack a crossbedded sandstone component but may contain a very fine, laminated or rippled sandstone at the base. The mud and silt components contain abundant fine plant material, including leaf impressions, but it is in the lignites that the most spectacular preservation of leaves, cones and seeds is found. The principal exhumed fossil forest sites are contained within this facies association.

Fossil tree stumps in growth and prone position are common, and many of the thick leaf litters are found around these original growth sites. Many upright stumps are hollow and are filled with shale and plant debris. Fossil tree trunks, branches and twigs along bedding in the lignite seams display highly variable degrees of compression, ranging from as little as 2:1, to 10:1 (ratio of cross-section diameters). Opaque resin is common, both as irregular lumps (usually oxidized) and in small "teardrop" shapes.

Phytolitic layers

Distinctive, white weathering horizons are present underneath several lignite seams at the fossil forest site. They appear to be confined to a stratigraphic interval between 65 and 95 m in Section 28-85 (Fig. 19). Detailed analysis of the white layers by Tarnocai et al. (*this volume*), from which the

Figure 11. *An example of interbedded trough and planar-tabular crossbedded sandstone facies, and planar-tabular crossbedded conglomerate facies in the conglomerate-sandstone member. Note the scouring at the base of some conglomerate units. (Section 11-87, Fig. 2a.) The exposure is about 5 m high.*

following comments are extracted, indicates a very high quartz content (up to 98 per cent), wherein individual crystals have an unusual platy morphology. Quartz grain size distribution is quite distinct from that of the surrounding host rocks. None of the white layers is associated with B horizons, which are typical of many soil-forming processes.

Paleosols

The most prominent paleosols in the siltstone-coal facies association include Gleysols, consisting of mottled and loamy sand with some leached organic-rich layers, and Organic paleosols of lignitic peat (Tarnocai and Smith, *this volume*). Some permineralization of plant material is evident.

PALEOCURRENTS

General observations of clast imbrication in the conglomerates suggest that flow was toward the east and southeast. Crossbed measurements were taken in the sandstone facies, mostly in the sandstone-mudstone association (Fig. 6). The vector means at four stations range from 144 to 150 degrees; the combined vector mean is 146.5 degrees. Data collected by Bustin (1982) from similar areas show a slightly more easterly trend. No correction for tectonic dip is required.

ANALYSIS OF THE COALS

Based on vitrinite reflectance, the coals in the fossil forest succession can be classified as lignites (Ro varies between

Figure 12. *Fining-upward sandstone-coal (lignite) units in the upper coal member (large arrows at the top of lignites). Fining-upward unit (a) shows large accretionary foreset bedding of sandstone (small arrows) dipping to the left, whereas unit (b) contains abundant crossbedded sandstone and numerous channel discordances. Details of the thick, central lignite seams are shown in Figure 14. The cliff section is indicated by a solid triangle in Figure 15.*

0.14 and 0.47; Goodarzi et al., *this volume*). Maceral analysis of coals, mostly from the fining-upward siltstone-coal association, indicates that they are huminite-rich and inertinite-poor. Analysis of coal samples from the forest site also indicates that the boron content is consistently less than 40 ppm.

THE FOSSIL FOREST: VIEWS OF THE PAST

The paleoenvironmental scenario for the fossil forests depicts an alluvial fan, or series of coalescing fans and successive, downslope braidplains and meanderplains that prograded toward the southeast, fed by detritus eroded from the rising Princess Margaret Arch. A variety of depositional settings within this alluvial complex are inferred from the facies associations (summarized in Table 1). It is emphasized that while these associations provide a convenient means for description and interpretation of sedimentary environments, they are part of a broad depositional continuum.

The most proximal alluvial facies preserved in the Geodetic Hills region are found in the matrix-supported and thick conglomerate associations. Sedimentological aspects of the former are consistent with those found in debris flows (Steel et al., 1977; Nemec and Steel, 1984; Rust and Koster, 1984). The thick tabular bedded conglomerate units, on the other hand, were deposited under stream flow conditions. Intercalation of both of these associations, their relatively coarse nature compared with the other lithofacies, the moderate degree of rounding throughout, and the paucity of transverse-flow bedforms and fine grained facies, suggest that deposition took place on a middle to outer fan setting. Mechanisms of gravel transport were debris flow and sheet floods, the latter producing some longitudinal and sheet bars. The requirement of a muddy matrix for maintenance of viscosity and shear strength in debris flows (e.g., Middleton and Hampton, 1976) is satisfied in terms of source terrane, where thick Mesozoic shale successions on Princess Margaret Arch

Figure 13. *Profile of a fining-upward sandstone-coal unit showing details of the internal organization. Note the decreasing size of bedforms concomitant with the fining-upward trend. Bedding units with shallow inclination toward the left (arrows) are thought to be the remnants of a large accretionary foreset (cf. epsilon crossbedding or inclined heterolithic stratification). An exhumed fossil forest can be seen immediately behind and to the left of the exposure (Section 10-87.) Scale is 26 cm long.*

are intruded by abundant diabase sills. Oversized boulders at the base of a few tabular bedded conglomerate units (rare examples up to 50 cm across) may represent stream reworking of debris flow material. This interplay between sediment gravity flow and bedload mechanisms reflects the ephemeral nature of the outer fan/braidplain transition that shifts according to the changing sediment flux. Only rarely did debris flows travel east of the present position of the Stolz Thrust, as indicated by examples of the matrix-supported facies.

Many of the features displayed by the conglomerate-sandstone association are consistent with an interpretation as a braidplain deposit. Unlike braided trunk rivers, braidplains merge upstream into alluvial fans (Rust, 1984). Crudely stratified and tabular crossbedded conglomerate facies compare favourably with modern analogues of longitudinal and transverse bars, respectively (see Hein and Walker, 1977). Many of the bars that formed during primary flood stages of stream flow were tens of metres long. Thin sandstone drapes and wedges that cap the gravel bars probably accumulated during the waning stages of flow (Rust, 1972, 1984), as bar tops were exposed and sediment transport was diverted into local bar-top channels and bar edges. On the gravel dominated parts of the braidplain, there is no evidence of overbank development, either adjacent to channel margins or on bar tops. Also, there are no fining- or coarsening-upward trends on the scale of individual channels and it is therefore likely that channel positions were markedly ephemeral. However, larger scale successions (50 to 80 m) do show general fining-upward trends, where mid-channel sand bars replace some of the higher energy, upstream gravel equivalents. Rather than reflecting the behaviour of single channels, the scale of these successions represents a trend to more distal (*sensu lato*) braidplain settings, either as the response to a retreating source, or to redirection of coarse sediment transport to another area of the plain.

Figure 14. *Details of a lignite seam, 1.6 m thick, capping a fining-upward sequence, corresponding to the middle seam in Figure 12. Contact with the subjacent siltstone (immediately above pen) is gradational. Note the alternation of papery, woody lignite layers (L) with mud-rich layers (M).*

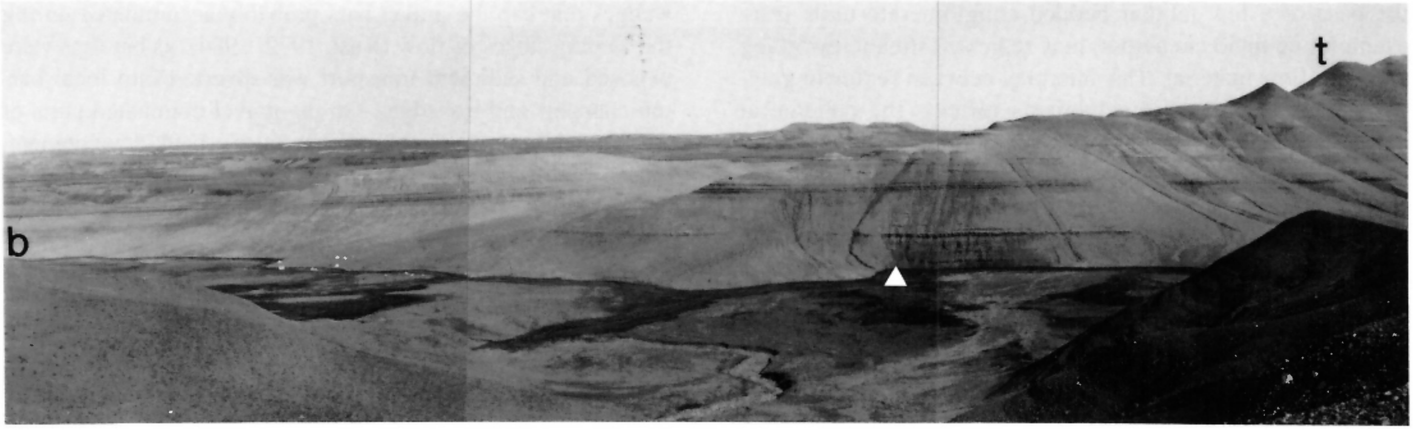

Figure 15. *Panorama of the fossil forest site showing the greatest exposed thickness of the upper coal member. Tents for scale at left-centre. The approximate base of Section 28-85 (Fig. 6) is indicated by "b" and the top by "t". Solid triangle indicates exposures in Figures 12 and 14.*

Figure 16. *A thick, trough crossbedded sandstone fining upward to a muddy lignite (at the top of the photo), exposed at the 65 m level in Section 28-85 (Fig. 2a). The major discordance (arrows) marks the base of an exhumed channel that has a maximum thickness of 4 m. The scale (cobblestone inside circle) is about 25 cm in diameter.*

Figure 17. An exhumed channel floor littered with tree trunks and other plant debris, from the 57.5 m level of Section 28-85 (Fig. 2a). The fossil tree trunk at left-centre is about 3 m long.

Both the thick sandstone and the conglomerate-mudstone associations can be viewed as facies that are transitional between the braidplain and meanderplain. The former, lacking conglomerate, was probably remote, either distally or laterally, from major coarse-sediment supply routes. High sinuosity streams can reasonably be discounted for the thick sandstone association because of the absence of fining-upward sequences and significant mud components. Most of the bedforms are consistent with a sandy braided environment (e.g., Miall, 1978), although the apparent stacking of channels could indicate an anastomosing system (e.g., Rust et al., 1984). Thin lignites and mudstones, which might represent wetlands and crevasse splays adjacent to anastomosing streams (Smith, 1983), are rare components of these associations. Indeed, the presence of some brown Podzolic horizons indicates some pedogenic processes, albeit in moderately well drained settings. But, as Rust (1981) has demonstrated for more arid settings, such facies are not a necessary prerequisite

Figure 18. Large, right-dipping accretionary foreset bedding (epsilon crossbedding) exposed in the cut bank along Section 11-87, upper coal member. Arrows locate the top and bottom of unit. Foreset is 3.5 to 4 m thick, and in this view extends laterally about 50 m. The unit is underlain by fining-upward sandstone-coal units and is overlain by thin fining-upward siltstone-coal units.

for development of anastomosing systems. No clear distinction can be made here between the sandy braided and anastomosing stream hypotheses for the thick sandstone association.

The conglomerate-mudstone association is interpreted as the product of gravelly meandering streams, based on the distinctive sequential changes in bedform and the consistently high proportion of muddy lithotypes (20 to 30 per cent) compared to most gravelly braided river facies. Modern counterparts of this facies association have been described from several places (e.g., McGowen and Garner, 1970; Bluck, 1971; Hooke, 1986). The conglomerate component might have accumulated at the apices of point bars at times of flood; both the upstream and downstream positions of the point bar would tend to develop interstratified sand and gravel. Different proportions of conglomerate in any individual fining-upward unit might then reflect positions on the point bar with respect to distance from the bar apex. However, it is possible that the conglomerates represent pool scour deposits (especially those having trough crossbedding) or mid-channel gravel bars. The sandier components of each unit accumulated on higher parts of the point bars. Local mud-filled channels that are cut into the sandstone (with gravel lags and climbing ripples) are similar to chute-fill deposits that develop on top of point bars during peak floods (e.g., McGowen and Garner, 1970). These gravelly meandering rivers were transitional between the braidplain and sand-dominated meanderplain, perhaps corresponding to the mixed-load channels of Schumm (1981), responding to marginally higher slopes and flood velocities compared to the downstream meander reaches.

Successions made up of sandstone-coal and siltstone-coal facies associations are the products of sandy meandering river channel and adjacent crevasse splay-floodbasin deposits. This interpretation is preferred over that employed for anastomosing river systems because of the consistent sequential changes

Figure 19. *A view of a (weathered) paleosol, including a distinctive, white weathering phytolitic bed, underlying a stump-bearing lignite, immediately below the 65.5 m level (Section 28-85, Fig. 6.) This horizon is truncated by the succeeding channel-fill sandstone depicted in Figure 16. Scale (notebook in circle) is 26 cm long.*

in bedform and grain size, the high proportion of fine grained lithotypes, the presence of epsilon bedding typical of point bars, and the absence of multistoried channel deposits. For the point bars in particular, criteria such as the predominance of sandstone, the absence of regularly intercalated sand-mud couplets and bioturbation, and the lack of other bedforms indicative of tidal influences (e.g., Smith, 1988) suggest that an estuarine setting is unlikely. Low boron contents in the fossil forest coals also suggests fresh water conditions. The fining-upward successions (Allen, 1965; Plint, 1983) suggest gradual migration of channels, where the siltstone-mudstone lithotypes indicate gradually encroaching floodbasin deposits over the point bars which ultimately are overlain by mixed conifer and hardwood forest. At the fossil forest site itself, the siltstone-coal associations are intimately associated with the meander channel and point bar deposits, but are much thinner and more numerous. They represent successive crevasse splays and more distal floodbasin accumulations during periods of flooding. Those fining-upward units containing basal, very fine, rippled sandstone are the more proximal splay deposits. Conifer and hardwood tree stumps in many of the lignites indicate climaxes in vegetation cover following each pulse of flooding. Evidence of minor flooding is also seen in the lignites in the form of thin mudstone interbeds.

At any particular time, much of the meanderplain must have been covered with forests at various stages of floral evolution. Only during major flood events, with extensive crevassing or channel avulsion, were forest conditions terminated over wide areas. The fining-upward sandstone-coal association, characterized by Podzolic and Gleyed Podzol paleosols was probably a better drained setting than its more distal counterpart, the siltstone-coal association. In the latter, Gleysols predominate, indicating poorer drainage, especially when associated with Organic (peaty) paleosols. Thus, although the floodplain setting envisaged for the siltstone-coal association is more distal to the main active meander channels, it appears to have been inundated for longer periods of time. Supporting evidence for prolonged "wet" conditions is found in the huminite-dominated coal macerals and in the excellent preservation of plant cellular material that requires protection from oxidation and decay. The paucity of inertinite suggests that swamp fires were infrequent, at least in the vicinity of the fossil forest site.

For certain flood events, it is clear that partial decay of tree stumps and litter had occurred before burial by sand or mud. Prolonged high water-table levels resulted in "heart rot" in many of the large tree trunks, where the inner trunk was removed and filled with carbonaceous shale and leaves. Variable amounts of compression in prone tree trunks, branches and twigs may also reflect different degrees of rotting prior to peat burial.

Each paleosol formed during the successive stages of forestation and floodplain aggradation. Presumably, each Podzolic paleosol had a laterally equivalent Gleysol on the adjacent floodplain, although such a transition has yet to be traced out in the field. Distinction between raised and low-lying peat swamps (e.g., Pocknall and Flores, 1987) may be possible, given these two "end-member" paleosol types. Peats developed in the fining-upward sandstone-coal association, comprising well drained channel and channel margin (perhaps levee) facies, were likely raised relative to the local water table. Likewise, peats associated with the more distal, poorly drained, and frequently wet floodplain Gleysols would have been low-lying.

Only approximate dimensions of the meander channels can be determined. Small channels were probably 30 to 50 m wide (based on very few preserved cross-sections); many were probably larger. Some of the smaller channel structures may in fact be chutes established on much larger channel-point bar systems. Estimates of water depths are always problematic, given that many of the channel fill sequences are truncated. However, the point bar deposits can be taken as a guide to approximate water depths of 2 to 4 m during periods of flooding. Given that the paleoslope in this area dipped southeast and that the continuity of individual sandstone-lignite beds in a northeast direction is at least 5 km (i.e., normal to stream flow), the meander belt at this site also had a minimum width of 5 km. However, meander belt widths must have varied considerably along the length of the alluvial plain because differential uplift and elevation of the source terrane along tectonic strike probably led to different rates of sediment supply and alluvial plain aggradation.

The directions of sediment transport between the braidplain (transport toward the east-southeast) and meanderplain (a southeast trend) in the Geodetic Hills area suggest a change in stream paleoflow outboard of the Eurekan mountain chain. Notably, the alluvial plain drainage appears to parallel tectonic strike and approximately parallels the basin axis. It is not known whether this trend persists in other areas of the synorogenic basin, or even if the trend was consistent throughout its history.

SOME CONTROLS ON THE HIERARCHY OF FOREST-FORMING ENVIRONMENTS

Colonization of the alluvial plain by plants was temporally and spatially varied, as indicated by the stratigraphic distribution of different facies associations that contain lignite seams or plant debris. Woody and leafy debris is dispersed throughout the conglomerate-bearing facies and, although the alluvial fan and braidplain was vegetated at least sporadically, widespread forest conditions did not develop in these less than hospitable settings. Only on the meanderplain, where interchannel areas were more stable, did colonization culminate in widespread mixed conifer and hardwood forests.

The distribution of lithofacies in space and time, including plant-bearing facies, is determined by both extrinsic (external to the sedimentary basin) and intrinsic (within the basin and depositional realm) variables. These factors, which have been discussed by Beerbower (1964) and Schumm (1981), can be evaluated in terms of stratigraphic sequences of different scales. At any time, sequences at all scales were in operation. Examples of other alluvial basins evaluated in this way are given by Heward (1978) and Ricketts (1986b).

Identification of depositional cycles is relatively straightforward and in this study is based almost entirely on field observations. As is the case in most sedimentary basins, however, the link between cause and effect of cyclicity is far more difficult to explain; the relative contributions of processes such as tectonics, isostasy, subsidence, sediment supply and changes in eustatic sea level are difficult to decipher – for some almost impossible (e.g., Burton et al., 1987). For example, even subtle changes in the position of the strandline will affect stream base levels and therefore the distribution of multifarious terrestrial facies. Specifically for the Buchanan Lake Formation on Axel Heiberg Island, no direct connection to the sea has been found, either in its sedimentary facies or boron content of the coals. Thus, the following tentative position is taken with regard to assigning causes to the observed cycles:

1. The Buchanan Lake Formation is a direct response to a major stage of Eurekan tectonism.
2. The formation on eastern Axel Heiberg Island accumulated in a narrow intermontane basin.
3. The principal controls on sedimentation for the major cycles were tectonic, overshadowing any eustatic effects. Eustatism, however, may have played a role in development of the minor cycles. (But, given the absence of marine-influenced facies, it may not be possible to test this latter hypothesis.)

The stratigraphic sequences need to be viewed in terms of the overall framework of the Axel Heiberg synorogenic basin. Sediment was supplied from Mesozoic and upper Paleozoic rocks being uplifted, folded, and faulted in central and western Axel Heiberg Island (Princess Margaret Arch). These events produced the Buchanan Lake Formation strata. Over most of eastern Axel Heiberg Island, Buchanan Lake rocks overlie folded and faulted Mesozoic strata with profound unconformity. The preserved sedimentary record suggests that these events are confined to the medial Eocene, although they may have extended into the Late Eocene and even earliest Oligocene. Furthermore, the youngest beds in the Iceberg Bay Formation, which disconformably underlies the Buchanan Lake Formation, are also Middle Eocene in age. Thus, the main period of deformation on Axel Heiberg Island is also constrained to the Middle Eocene and presumably occurred during the hiatus represented by the Iceberg Bay-Buchanan Lake unconformity. The Buchanan Lake Formation itself therefore represents a relatively late stage in the Eurekan tectonic record. Sediment deposited during the early stage of Eurekan deformation is no longer preserved onshore (perhaps it was cannibalized), but may be present beneath the polar continental shelf farther west.

Stratigraphic sequences identified for the fossil forest succession are summarized in Figure 20. First- and second-order sequences are controlled by extrinsic factors, whereas third- and fourth-order sequences are controlled by intrinsic factors. First-order sequences several hundred metres thick reflect major periods of sediment supply in response to uplift, folding, and faulting that affected the entire basin. They may also provide evidence of isostatic adjustment to the downward flexure of the basin caused by tectonic loading in the adjacent mountains. Two, and possibly three, major pulses of sediment influx occurred in the Geodetic Hills area. The lower coal member and its possible correlative, the conglomerate member, provide the earliest sedimentological evidence of major Eurekan tectonism. The second pulse is delimited by the conglomerate-sandstone and upper coal members. Approximately 100 m of sandstone-dominated beds at the top of the fossil forest succession (Section 28-85) may indicate a third important change in sediment supply, but this event is poorly constrained. Each first-order sequence exhibits overall fining-upward trends, reflecting laterally equivalent outermost fan, braidplain, and meanderplain transitions and gradually decreasing sediment supply, presumably by erosion and reduction of relief. The transition from one first-order sequence to another indicates a major period of renewed uplift.

Second-order sequences, which also show broad fining-upward trends, are 50 to 80 m thick in the conglomerate-sandstone member and 20 to 30 m thick in the upper coal member. Both mimic the first-order trends in that they represent increasingly distal environments on braidplains and meanderplains. For example, in the conglomerate-sandstone member, the trends are indicative of downslope changes from gravel dominated channels to mixed sand and gravel channels. Second-order transitions in the upper coal member, on the other hand, consist of meander channel sands to successive, distal floodbasin deposits. The second-order trends probably formed in response to more localized, lower amplitude tectonic events and subsequent erosion and source retreat than those envisaged for the first-order sequences. Some response to isostatic adjustment in the basin may also be represented but distinction between isostatic and tectonic processes is difficult. This type of sequence may correspond in part to the "geomorphic threshold" discussed by Schumm (1981).

Processes controlling the third- and fourth-order sequences were probably intrinsic to the sedimentary system, although minor shifts in eustatic sea level may have played some role. Fining-upward sequences, generally less than 10 to 12 m thick, are in many cases equivalent to specific facies associations. For example, on the meanderplain such trends have been inferred to represent laterally migrating meander channels. Third-order transitions, from floodplain deposits to channel sands, probably occurred in response to local gradient advantages on the aggrading alluvial plain, to periodic floods, or (as Smith, 1983, has suggested) by avulsion resulting from log-jams.

Modern Podzolic and Gleysolic soils found on meanderplain channel margins and floodplains in humid temperate climates form in time frames of 1000 to 10 000 years (Tarnocai et al., *this volume*). Thus, by analogy, the fossil forest paleosols represent minimum time limits for accumulation of the third-order, fining-upward sandstone-coal and siltstone-coal sequences (N.B., this only represents the time for soil accumulation and not each complete fining-upward sequence). What is not discernible, however, is the duration of possible hiatuses between successive fining-upward sequences. Over a stratigraphic interval of about 90 m in Section 28-85, there are at least 50 coal/paleosols, many of which are composite. Thus, for the paleosol component alone, the time of accumulation could be 0.5 million years – the stratigraphic interval here representing about 30 per cent of the upper coal member and about 10 per cent of the total Buchanan Lake Formation at Geodetic Hills.

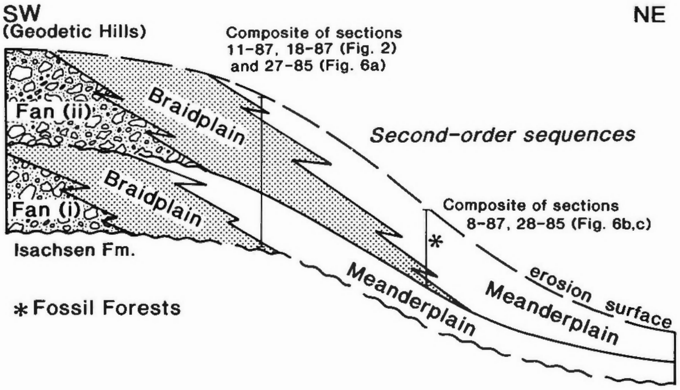

Figure 20. *A schematic representation of interpreted first- and second-order sequence relationships and unconformities in the Buchanan Lake Formation, Geodetic Hills area. First-order sequences are represented by fans (i) and (ii). The locations of measured sections indicate positions during the main phase of Eurekan deformation; i.e., during uplift and erosion of ancestral Princess Margaret Arch, but prior to "final" movement on the Stolz frontal thrust. Note that the alluvial fan component of sequence (ii) has been almost completely eroded but its hypothetical position is shown here; rare debris flow breccias in the conglomerate-sandstone member may represent the distal edge of this alluvial fan.*

Fourth-order sequences occur on a scale of single beds or groups of beds, ranging in thickness from centimetres to decimetres. Many, such as mudstone partings in lignite seams and thin crevasse splay sandstones, represent single depositional events.

SUMMARY

Folding and faulting during the major phase of Eurekan tectonism fragmented older Eureka Sound Group basins into narrow, intermontane troughs or basins. Near Geodetic Hills, the site of the fossil forest, coarse sediment was shed off the uplifted terrane in two, or possibly three, pulses, each related to successive uplifts and source retreat (erosion). Sediment eroded from the uplift was dispersed onto an alluvial plain by a variety of drainage types, depending on: the proximity to the source (which also changed with time), the relative effects of subsidence caused by tectonic loading and isostatic adjustment to erosion, regional slope, and more local effects intrinsic to the depositional system, for example, gradient advantages and channel switching.

The following facies domains are recognized (Fig. 21):

1. Middle and outer alluvial fan (matrix-supported conglomerate association), possibly transitional to proximal braidplain (thick conglomerate association), including debris-flow breccias.
2. Proximal braidplain, characterized by longitudinal and transverse gravel bars (conglomerate-sandstone association).
3. Two facies associations that represent the transition from braidplain to meanderplain. The first is a thick sandstone association for which the choice between sandy braided or anastomosing channel deposits is equivocal. The second transitional facies includes a conglomerate-mudstone association, which is considered to be indicative of gravelly meandering stream deposition.
4. Facies indicative of lowest slope and stream competence on the alluvial plain include sandy meandering rivers and overbank deposits (sandstone-coal association). Point bars and chute channels are exposed locally. Channel widths were probably many tens of metres, and meander belt widths were 5 km and more in the study area.
5. Closely linked with the meander channel sequences are siltstone-coal associations that provide evidence

for extensive floodplain and crevasse splay deposition and development of mixed conifer-hardwood forests at more than 50 stratigraphic levels (Section 28-85).

Stratigraphic sequences, all of which fine upward, can be viewed as stratigraphic building blocks that reflect basic sedimentological and tectonic processes of basin fill. Four sequence orders are recognized:

1. First-order sequences reflect major tectonic events that probably affected all or large parts of the basin, in terms of uplift, erosion, and sediment supply. Figure 20 illustrates the inferred sequence relationships before tectonic shortening along the (frontal) Stolz Fault zone.
2. Second-order events represent more localized tectonism and perhaps isostatic adjustments in response to tectonic loading.
3. Accumulation of third-order fining-upward sequences reflects processes intrinsic to the sedimentary system (for example, major flood events, or channel switching because of local gradient advantages), although subtle eustatic effects may have played some role. Based on paleosol types, the third-order sequences represent accumulation periods of at least 10 000 years.
4. Fourth-order sequences are probably the closest approximation to single depositional events commonly represented by single beds.

The general scenario of Buchanan Lake Formation deposition at the fossil forest site and environs is one of a sparsely vegetated, gravelly braidplain fed by alluvial fans, grading downslope to a broad meanderplain that was drained by high sinuosity rivers adjacent to densely vegetated floodplains (Fig. 21). Stratigraphic organization of the basin fill indicates continual migration of these environments in relation to the source terrane. Preliminary structural considerations (Ricketts, 1987b) suggest up to 10 km of structural shortening across the Stolz Thrust. The fossil forest site is about 8 km northeast of the main thrust. Therefore, a distance of 10 to 18 km can be regarded as an approximate measure for the separation of the forest sites from the ancestral Princess Margaret Mountains.

As a cautionary note, however, it should be pointed out that the measure of structural shortening is the cumulative effect of all Eurekan shortening events and hence may be overestimated for the specific stratigraphic levels in which the fossil forests occur.

Progradation of the alluvial fans and braidplain was east-southeast. Stream flow on the meanderplain appears to have been diverted to a more consistent southeast direction, almost parallel to tectonic strike. This trend may reflect a regional,

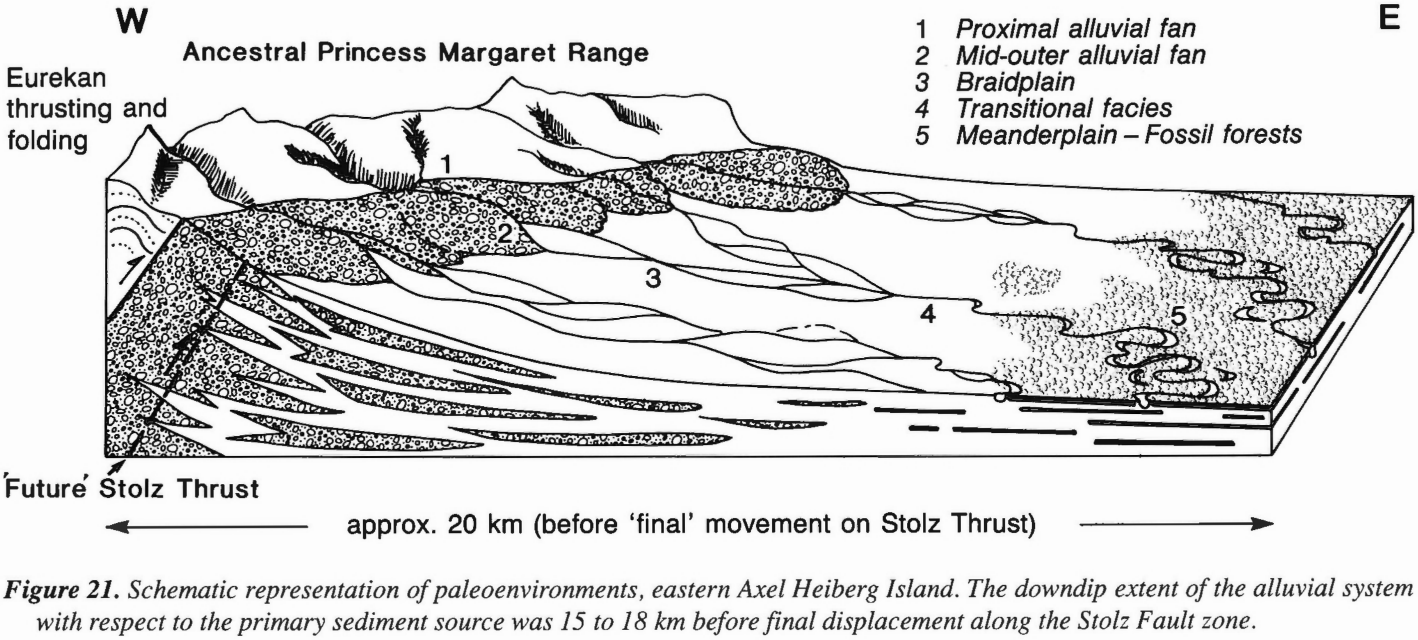

Figure 21. *Schematic representation of paleoenvironments, eastern Axel Heiberg Island. The downdip extent of the alluvial system with respect to the primary sediment source was 15 to 18 km before final displacement along the Stolz Fault zone.*

southeast dipping basin axis, with the basin being closed to the northwest, and opening toward the south; i.e., the meanderplain component represents the transition to axial drainage.

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THE DYNAMICS OF POLAR FOSSIL FORESTS: TERTIARY FOSSIL FORESTS OF AXEL HEIBERG ISLAND, CANADIAN ARCTIC ARCHIPELAGO

Jane E. Francis[1]

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Abstract

Fossil forests are preserved within sediments of Middle Eocene age at Geodetic Hills, eastern Axel Heiberg Island. They represent some of the most high-latitude forests ever reported. Layers of mummified leaf litter, cones and seeds, plus large in situ fossil tree stumps are interbedded with paleosols, siltstones and sandstones of the Eureka Sound Group. They represent an environment of floodplains adjacent to a large river system, over which dense forests of deciduous conifers (principally *Metasequoia*) and broad-leaved angiosperms became established intermittently.

Several hundred tree stumps are exposed in over 30 forest layers at this locality. At two particular levels, large areas of fossil forests have been exposed, revealing numerous large stumps and horizontal logs. The densities of the stumps for these two areas are: 1 stump in 21 m^2 (484 a hectare) and 1 stump in 31 m^2 (325 a hectare). The stumps are up to 0.9 m in height and 3.75 m in diameter (diameter of root stock) and vary from conical to dome-shaped.

Fossil forests at such high latitudes represent a unique environment not present on earth today. Wide growth rings in the fossil wood illustrate that ambient temperatures were warm and favourable for forest growth, and that trees were adapted to the extended periods of daylight and darkness and the predominantly low-angle sunlight present at that time. Organic productivity was high, estimated to be approximately 1200 $gm/m^2/y$, and comparable to the productivity of living temperate, deciduous forests.

Résumé

Des forêts fossiles sont conservées dans les sédiments de l'Éocène moyen à l'emplacement des collines Geodetic, dans l'est de l'île Axel Heiberg. Elles représentent quelques-unes des forêts les plus septentrionales jamais signalées. Des couches de litière feuillue, de cônes et de graines momifiés, ainsi que de grosses souches d'arbres fossiles trouvées in situ, sont interstratifiés avec des paléosols, des siltstones et des grès du groupe d'Eureka Sound. Ils représentent un milieu de plaines inondables proches d'un vaste réseau hydrographique, sur lesquelles de denses forêts de conifères à feuilles caduques (principalement *Metasequoia*) et d'angiospermes à grandes feuilles se sont établies de façon intermittente.

Plusieurs centaines de souches d'arbres affleurent dans plus de 30 couches forestières dans cette localité. À deux niveaux particuliers, de grandes zones de forêts fossiles sont exposées, et de grosses souches, ainsi que de grands troncs disposés à l'horizontale sont en évidence. Dans ces deux régions, les densités des souches sont les suivantes: une souche pour 21 m^2 (484 souches par hectare) et une souche pour 31 m^2 (325 par hectare). Les souches ont jusqu'à 0,9 m de haut et 3,75 m de diamètre (diamètre du rhizome), et leur forme varie de celle d'un cône à celle d'un dôme.

A des latitudes aussi élevées, les forêts fossiles représentent un milieu unique qui n'existe nulle part sur la terre de nos jours. Dans le bois fossile, les larges anneaux de croissance indiquent que les températures ambiantes étaient élevées et favorables au développement des forêts, et que les arbres étaient adaptés aux longues périodes de lumière du jour et d'obscurité, et à l'ensoleillement qui était principalement rasant à cette époque. La productivité organique était élevée, et selon les estimations d'environ 1200 $g/m^2/an$ et comparable à la productivité des forêts actuelles d'arbres à feuilles caduques des climats tempérés.

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[1] Department of Geology and Geophysics, University of Adelaide, P.O. Box 498, Adelaide, South Australia 5001, Australia

INTRODUCTION

The presence of forests of large trees in polar regions during the geological past indicates a unique environment for which there is no modern analogue. Trees are unable to grow at high latitudes today due to the low temperatures in the present glacial climate. The effect on trees of another influential parameter, the uneven annual distribution of sunlight, is not well known. Although it has been proposed that, theoretically, trees could have adapted to the extreme seasonality of the light regime, it is only from the study of the fossilized remains of polar forests that some idea can be obtained about how plants grew in the past under these specialized environmental conditions.

Important information can also be obtained for reconstructing past climates. The fossil forests existed near the pole when there was no polar ice cap and ambient temperatures were much higher (Frakes, 1979). The presence of other fossil plant assemblages of similar ages in regions of high latitude indicates that the Middle Eocene was one of the warmest intervals in the late Phanerozoic (Wolfe and Poore, 1982).

The forests harbour important evidence about environments and climates that no longer exist on Earth today. To understand the dynamics of forest growth in this unusual environment it is necessary to build an accurate reconstruction of the environment from the fossil evidence. This paper presents preliminary information from the spectacular, mummified forests at eastern Geodetic Hills on Axel Heiberg Island.

GEOLOGICAL SETTING

The fossil forests of eastern Axel Heiberg Island are well exposed on two northeast-trending ridges (Fig. 1). The forest layers are preserved as dark, lignitic horizons of leafy plant material and fossil wood, and are occasionally underlain by white clay horizons, which are the remains of paleosols. The organic layers are interbedded with poorly lithified siltstone and sandstone of the Buchanan Lake Formation, part of the Eureka Sound Group (Ricketts, 1986). The sediments represent floodbasin deposits, crevasse splays and high-sinuosity river sediments on a broad alluvial plain. The fining-upward sequences of river sands, floodplain silts and vegetated soils indicate a history of cyclical river flooding followed by the establishment of forest vegetation and, at times when flooding episodes were not frequent, the development of leached soils.

The sediments have been dated as Paleogene on the basis of pollen analysis. The flora is considered to be Middle Eocene in age, possibly extending to earliest Oligocene (McIntyre, *this volume*). The paleolatitude at this time was possibly as high as 78°N (McKenna, 1980).

THE FOSSIL FORESTS

Over 30 layers of fossil forests are present on the ridges at the Geodetic Hills site (Fig. 1). They are preserved as lignitic layers from a few centimetres in thickness, to beds over one metre thick. Fossil leaf litter is the principal component of the lignite, along with tree-root material, representing relatively

***Figure 1**. The fossil forest ridge at the Geodetic Hills locality, eastern Axel Heiberg, looking northeast. The dark bands are layers of mummified forest floor litter with tree stumps.*

undisturbed debris from forest floors. The state of preservation is quite remarkable in that the organic material is predominantly unconsolidated, particularly near the upper layers, and individual leaves and plant organs can be easily retrieved. There are only very minor amounts of sediment within the leaf litter layers and very little mineralized matter (though roots in a few zones have been replaced by siderite in some parts of the section). The lower layers of the lignite have undergone some compression and have the appearance of low-grade coals.

Much of the lignite contains large tree stumps in their original growth position. On certain ridge tops, large areas of lignite have been uncovered, revealing tree stumps among the leaf litter of the former forest floor. One layer on the southwest end of the forest ridge, level N (Fig. 2), contains large, woody stumps surrounded by leaf litter, and appears to represent relatively undisturbed forest floor conditions. In contrast, in another layer (layer HR) at the northeast end, upright stumps and horizontal tree trunks are preserved within silts and sands and the leaf litter layer is absent, presumably having been washed away from around the tree stumps prior to burial by the flood sediments (Fig. 3A). The horizontal logs probably represent trunks and branches that fell from the decaying stumps and were washed into this area from adjacent parts of the forest.

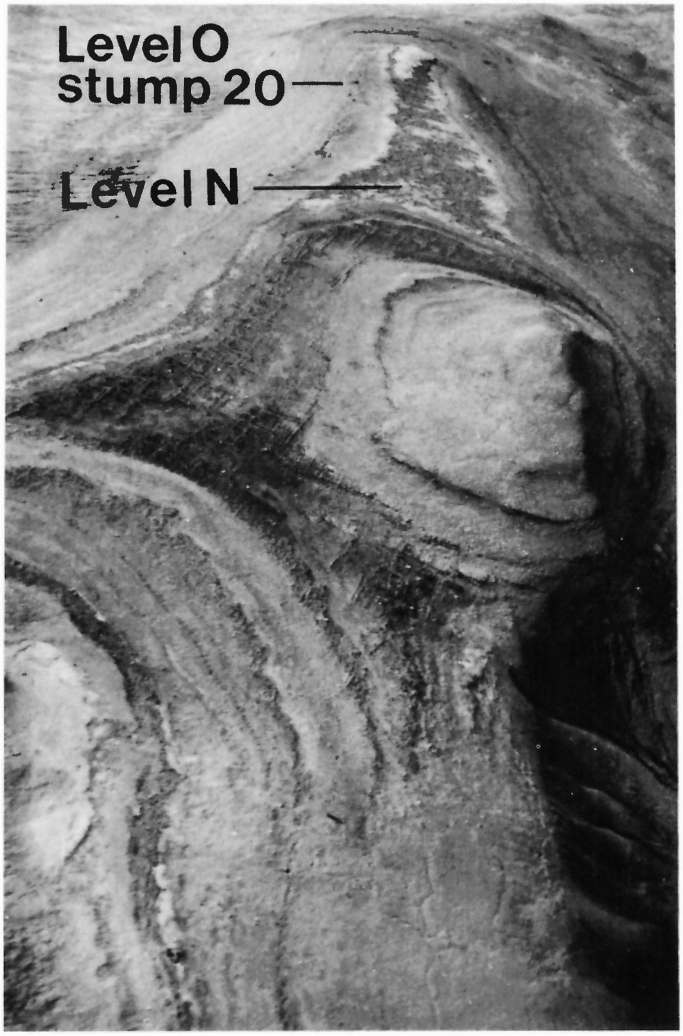

Figure 2. The fossil forest ridge from the air, looking down on the western end of the ridge. Forest floor litter layers are exposed on plateau surfaces. Levels N and O and the position of the large stump (20) at Level O are indicated (see Fig. 3E).

Many of the tree stumps of level HR have hollowed trunks filled with silt containing leaf impressions (Fig. 3B); these trunks must have been hollow during inundation by flood waters to allow silt and leaf debris to be washed in. The hollows may represent some kind of heart rot in old trees on the forest floor, or rotting caused by early stages of flooding and waterlogging.

The fossil wood in these forests is also exceptional in that it is not petrified but "mummified", and retains its woody nature. It has the appearance of blackened, recently living wood, burns easily and can be cut with a saw. It retains a large quantity of water and cracks rapidly upon exposure to the air.

From study of thin sections, it appears that the wood cells have collapsed during waterlogging and burial, perhaps due to the lack of permineralizing solutions that would have enabled the retention of the three-dimensional structure. Cellular structural detail has been lost so that detailed identification is not possible (Fig. 4B, C), and the wood has undergone the primary stages of gelification. In a few trees that are in part undistorted, growth rings 3 mm in width, and some up to 5 mm, have been observed, indicative of fast growth in a favourable climate (Fig. 4A).

Tree dimensions and shape

Several hundred fossil tree stumps are exposed at various levels over the ridges. Some are visible only at the exposed edges of the lignite layers, while others are present over large areas on plateau surfaces. The stumps range in size from approximately 20 cm in diameter to very large root stocks several metres in diameter. The exposure of the tree stumps is variable and often only the top of the trunk is visible in the sand layers, so that it is difficult to assess the overall range in stump size. Several of the larger stumps were excavated in order to investigate root characteristics.

The tree stumps among the leaf litter at level N are fully exposed and have been eroded down to the basal part of the root stock. The root systems have been vertically compressed into the leaf litter and their original shapes are now difficult to determine. The root stocks at this level range from less than

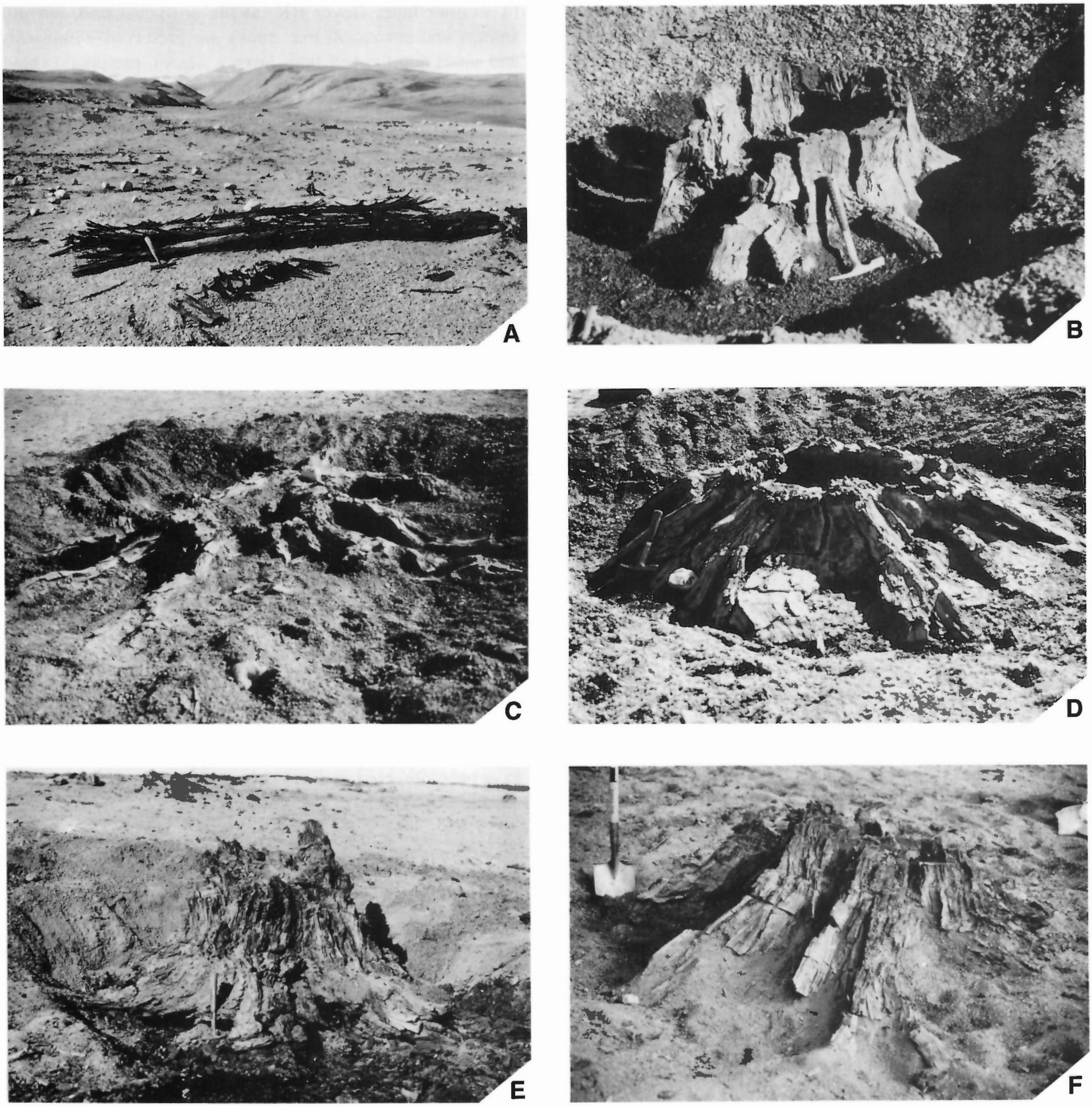

Figure 3*. Fossil wood from Geodetic Hills. A, large horizontal tree trunks on sands and silts of layer HR. The log in the foreground (log 13 in Fig. 6) is 10 m long; B, tree stump with hollowed rotted centre, now filled with siltstone containing leaf impressions. Level HR, no. 50. Stump diameter 1.10 m; C, stump with branching roots radiating out to a diameter of 3.75 m. Level HR, no. 21; D, dome-shaped stump with conical root stock. Level HR, no. 18. Stump diameter 2.10 m; E, large mummified stump at Level O. Height is 0.90 m, maximum diameter 1.70 m. Stump no. 20; F, mummified stump at Level O. Height 0.65 m, maximum diameter 1.45 m.*

10 cm to over 2 m in diameter. Upright parts of the trunks have not been preserved. However, in the underlying lignitic level (level O) large stumps have been preserved in a line along the edge of the ridge (Fig. 3E, F). The largest stump is 0.9 m in height, its trunk is 1 m in diameter and the roots extend to a diameter of 1.70 m (Fig. 3E). Bark is present in patches on the exterior of the trunk, a very rare occurrence for fossil trees. However, contortion of the wood grain on the upper surfaces of the stump indicates that the trunks have undergone both vertical and horizontal compression, so although the presence of bark on the exterior illustrates that the complete trunk is present, the original dimensions of the tree have been reduced by compression.

At the HR level the tree stumps in silts and sands have undergone less compaction and the shape of the root stock is still apparent (Fig. 3B-D). Excavation of several stumps revealed various root styles. Some roots are conical in shape, consisting of a dome over 2 m in diameter formed of a sheath of roots (Fig. 3D). Fossil tree stumps of very similar conical shape are present in another petrified forest, probably of Paleocene age, at Strathcona Fiord on Ellesmere Island (Francis, 1988). At that locality, all the stumps are of the same shape, suggesting a monotypic forest community. In contrast, another stump at the HR level had branched roots radiating horizontally from the centre to a maximum diameter of 3.75 m (Fig. 3C). A third, smaller tree had an upright trunk with very little of the root system revealed (Fig. 3B). Although the wood types of these different stumps have yet to be identified, the differing habit of the roots in such close proximity suggests that they belong to different types of trees.

Tree density

The density of a forest gives some indication of the limitations of environmental parameters, such as availability of water, light, soil stability, etc. Measurement of the density also provides quantification of what appears to have been crowded conditions in the ancient forests. The density of the stumps at both levels N and HR was assessed using a 5 m grid to construct a plan of each level, onto which the position of stumps, logs and the limits of exposure of the levels were plotted. The area of exposure was then measured, the number of stumps counted, and the density calculated.

At level N, 109 stumps were recorded in an area of 2250 m^2, giving a density of 1 stump in 21 m^2 (or 484 stumps a hectare) (Fig. 5). At level HR, 91 stumps were present in 2800 m^2, giving a density of one stump in 31 m^2 (or 325 stumps a hectare) (Fig. 6). In both cases the densities are probably

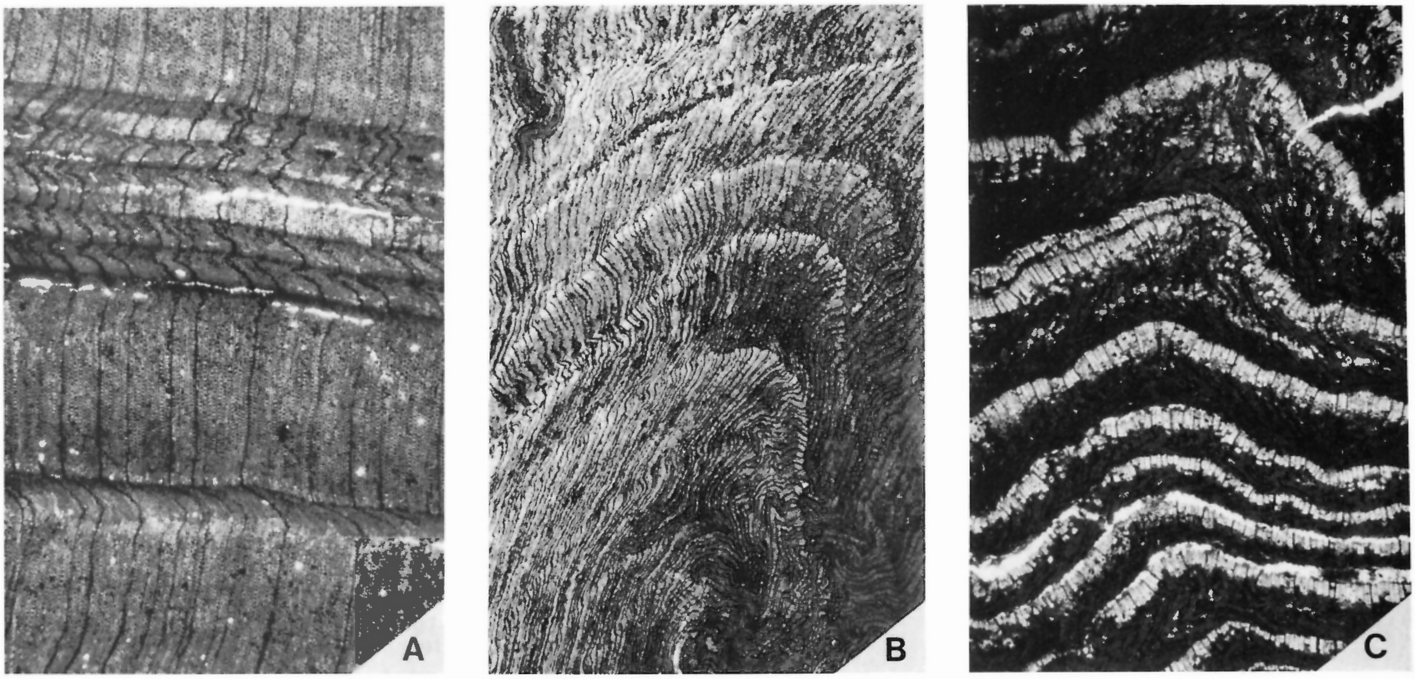

***Figure 4**. Traverse sections of fossil wood from Geodetic Hills fossil forest. Width of field of view 4 mm. A, GH 45, showing zones of well preserved and crushed rings; B, GH 43, showing contorted growth rings; C, GH 49, showing rings with both crushed and permineralized cells. The thick-walled cells of the latewood, formed at the end of the growing season, have been permineralized, but the larger, thin-walled cells of the earlywood zone have been compressed, prohibiting the infiltration of the mineral solution.*

underestimates because it is likely that other stumps are still buried under overlying sands, particularly at level HR.

The petrified forest at Strathcona Fiord, Ellesmere Island, has a stump density of one in 27 m^2, 367 a hectare; this order of magnitude therefore seems to be characteristic of these Tertiary high-latitude forests. As a comparison, modern cypress swamps in the Alabama wetlands, an analogue proposed for the Arctic forests, have tree densities of only 169 trees a hectare (Lugo, 1984). Densities in other natural modern forests range from, for example, 200-1000 trees a hectare in tropical rainforests, 1000-3000 a hectare in temperate broad-leaved evergreen forests, 485 a hectare in spruce forests in Finland, to over 3000 trees a hectare in boreal needle-leaved evergreen forests in Canada (Loucks et al., 1981).

Trunks and branches

Horizontal logs are a component of most of the forest layers. In the lignitic layers, however, the logs are compacted into the dark leaf litter and are hard to distinguish. In the silt and sand horizons, such as level HR, the logs retain their shape and are easily visible. They represent trunks and branches that have rotted from the tree stumps and have been transported by water (Fig. 3A). Sixty-two logs were measured, ranging in length from about 10 cm (branches less than 10 cm not recorded) to 11.5 m, and in width from a few centimetres to 70 cm. The surfaces of many of the logs have decayed, so the true trunk width is not apparent. The longest trunk, 11.5 m long, shows very little taper, suggesting that tree height was considerably greater than this. From the large size of the roots and trunks the average height of the trees is estimated to have been about 20 to 30 m.

The logs studied lie in two main directions: north-northeast (about 010°) and east-southeast (about 130°). These orientations may have resulted from storm damage. However, Ricketts (*this volume*) recorded similar paleocurrent directions to the east and southeast for the sediments. One set of fossil logs is consistent with this, suggesting reorientation by water currents even if they were not transported far. The other set of logs is orientated in an approximately perpendicular direction, perhaps representing logs caught on river bars or banks as log jams. One long log (No. 149 in Fig. 6) had obviously been jammed between tree trunks since it has three stumps in close proximity that would have prevented its movement.

FOREST DYNAMICS

Although forests do not now grow beyond about 70° latitude in the northern hemisphere and 56° in the south, forests near the poles were not uncommon in the past, particularly in the late Phanerozoic, as illustrated by reports of fossil forests from the Cretaceous of Antarctica (Jefferson, 1982; Francis,

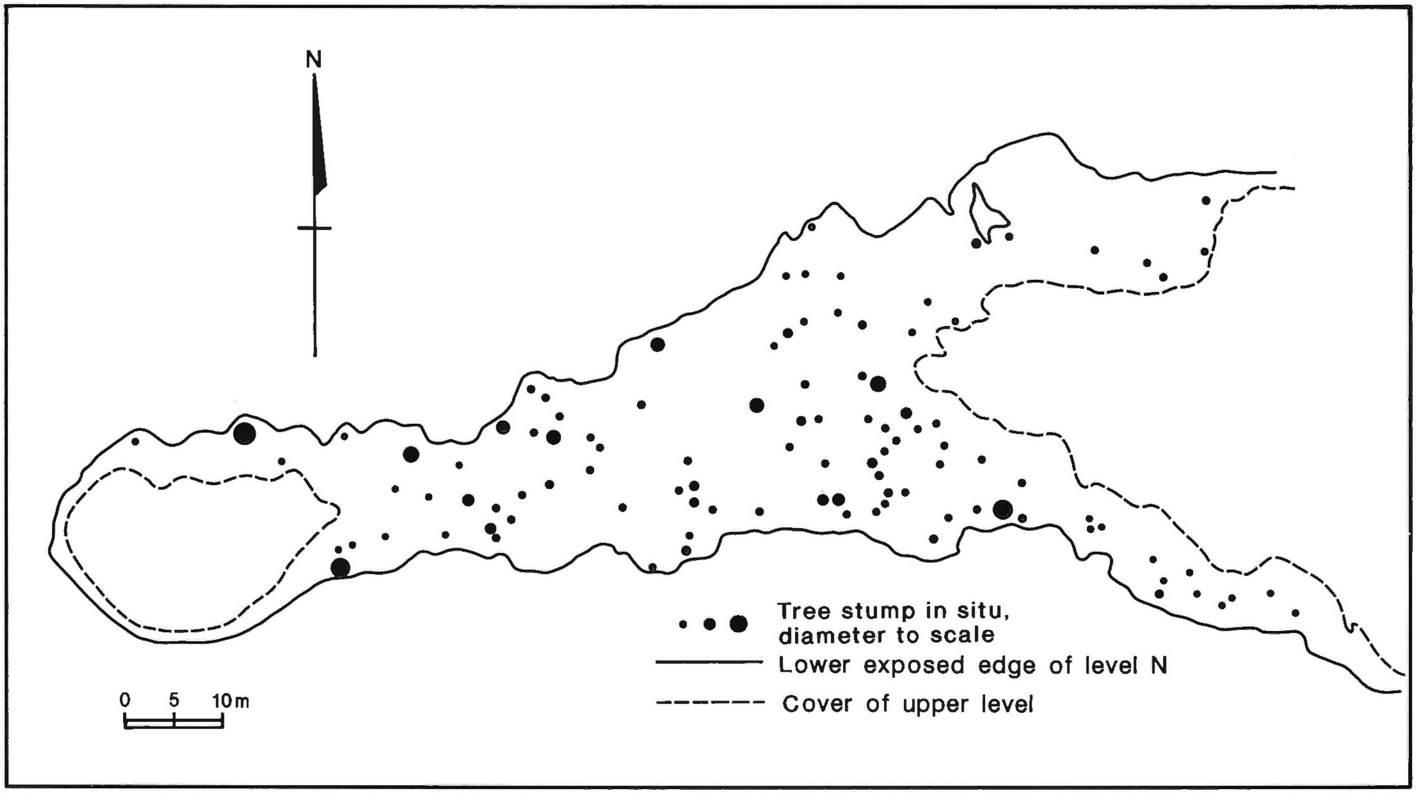

Figure 5. Plan view of level N showing distribution of in situ tree stumps. Diameters of the stumps are to scale.

1986), Australia (Douglas and Williams, 1982; Frakes and Francis, 1988), Alaska (Spicer and Parrish, 1986) and from the Tertiary of Spitsbergen (Schweitzer, 1980). The Eocene forests of Axel Heiberg are of special interest because of the exceptional preservation of so many environmental indicators.

From field observations reported here, a picture of the forests can be constructed. The dimensions of the stumps and logs illustrate that the trees were very large and certainly not small, shrubby bushes. Trees with monopodial trunks up to 1 m in diameter with estimated heights of 20 to 30 m are envisaged. If the observed ring widths of 3 mm are taken as the average record of growth, then trees of this size may have been over 150 years old or more.

Several types of trees grew in the forests, as indicated by the presence of a fairly diverse assemblage of fossil leaves in the forest litter. The most abundant foliage is that of the conifer *Metasequoia* (dawn redwood). The living trees of *Metasequoia*, once widespread in the northern hemisphere but now restricted to small communities in China (Chu and Cooper, 1950), are deciduous, which may account in part for the abundance of these leaves. The taxodiaceous conifer *Glyptostrobus* (swamp cypress) is also common (Basinger, 1986). Other leaf types in the litter include the conifers pine, larch (also deciduous) and fir, along with the angiosperms katsura (very common at some levels) and members of the birch, walnut and oak families (Basinger, 1986). Although the wood genera have yet to be identified, the different styles of roots in the fossil forests suggest that several types of tree are represented. Those of conical shape are very similar in appearance to living taxodiaceous swamp conifers (Lugo, 1984) and similar ones also occur in other arctic Tertiary forests, such as at Strathcona Fiord (Francis, 1988) and Stenkul Fiord (Francis, personal observation) on Ellesmere Island.

The existence of forests at high paleolatitudes has provoked much discussion because, even if the ambient

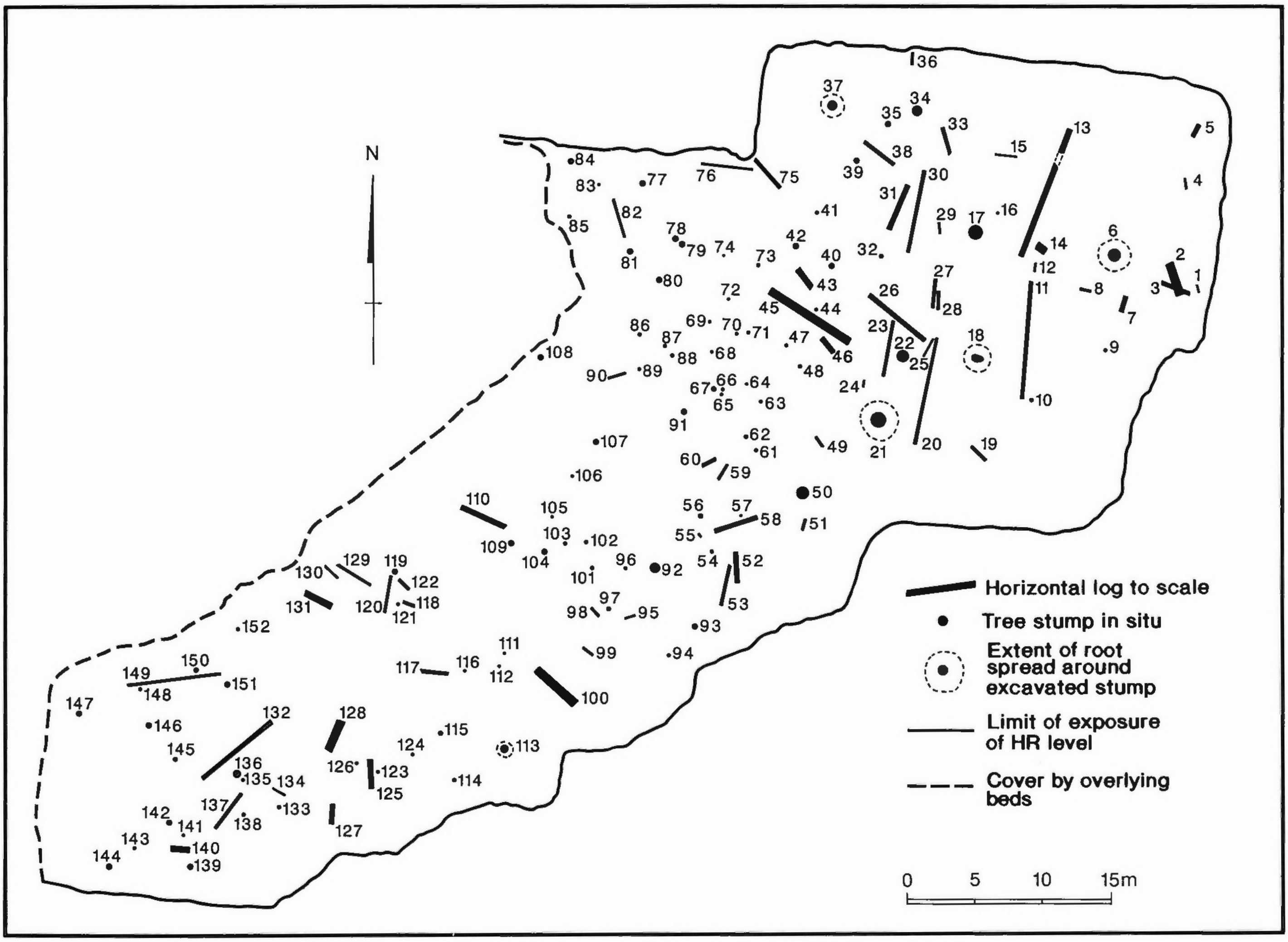

***Figure 6**. Plan view of level HR showing distribution of stumps and horizontal logs. The logs and stumps are to scale.*

temperatures had been high enough, these large forest trees would have had to grow in a regime of predominantly low-angle sunlight. A change in the axial tilt of the earth's rotational axis was proposed by Douglas and Williams (1982) and by Jefferson (1982) to provide a more even annual distribution of light, but climate modelling has shown that this would have resulted in lower polar temperatures (Barron, 1984) and less seasonal climates (Creber and Chaloner, 1985).

Despite the high latitude, the forest trees grew to large sizes and produced considerable quantities of wood. The climate was clearly suitable for forest growth and there was sufficient light, even though the plants had to tolerate a polar regime of months of winter darkness, intervening months of low-angle light and summer months of continuous sunlight. In fact, the total amount of sunlight in the polar regions in a year is equivalent to that received at lower latitudes because the sun shines for such long hours in summer (Creber and Chaloner, 1985). It appears that the trees were able to grow continuously during the warm summer season but probably became dormant during the cold dark winter months (Axelrod, 1984). The deciduous nature of many of the trees in the Axel Heiberg forests was probably an important mechanism that helped them to survive the dark winters, because the foliage could be shed to avoid food reserves depleting.

The density of trees in the fossil forests was fairly high, with 484 and 325 trees a hectare at the Geodetic Hills and 367 a hectare at Strathcona Fiord, approximately one tree in every 20 to 30 m^2. One of the environmental factors that might have limited tree density was the low-angle sunlight that would have been incident on the forest for many months of the year. Had the forest been too dense, the long shadows cast by tall trees would have severely affected the growth of adjacent trees. For this reason, Creber and Chaloner (1985), in their study of high-latitude fossil forests, suggested that the most suitable shape for polar trees to produce minimal mutual shading would have been a conical shape. Living trees growing near the northern treeline today have a thin, pencil-like habit. Living examples of *Metasequoia* and taxodiaceous conifers have a strikingly conical shape, and also have light, feathery foliage, which permits maximum penetration of light to lower branches; all these are useful characteristics for growing in polar light regimes.

An estimate of the production of woody material produced by a tree trunk from the Axel Heiberg forests can be made by considering the amount of wood produced in a growth ring in one year. If the trunk of a tree is regarded as a cone, the volume of wood produced in one year will be the difference between the volume of wood in the cone at the beginning of the growing season and that at the end (Creber and Francis, 1987). For example, if the average height of a tree in these fossil forests was 20 m, the average trunk base 50 cm, and the average ring width 3 mm, the added volume of wood in a year would be 0.063 m^3. Taking the gravimetric density of conifer wood to be 0.4 gm/cm^3 and the forest density as one tree in 21 m^2, then the yield of wood per year would be 1200 gm/m^2. This is, of course, a very approximate value but can be refined in future with more detailed analyses.

In comparison with living forest production, this value represents quite a high yield and is most comparable to that of modern temperate forests. For example, annual production in temperate deciduous forests reaches 1200±25 per cent $gm/m^2/y$ (Art and Marks, 1978), compared to 3250 to 5000 $gm/m^2/y$ in tropical rainforests, 800 $gm/m^2/y$ in boreal forests and only 20 to 130 $gm/m^2/y$ in tundra habitats on Devon Island, Canada (Wielgolaski, 1978). Annual production in a natural stand of balsam fir (*Abies balsamea*) in Canada was recorded as 935 to 1258 $gm/m^2/y$ (Art and Marks, 1978). The estimated production of wood material from the fossil forests therefore suggests that they are more similar to living temperate deciduous forests than tropical or boreal types.

The Tertiary was a period of global cooling, transitional from the warm climates of the Late Cretaceous to the onset of glaciation in Antarctica during the Oligocene. However, superimposed on this broad cooling trend were intervals of warming, including intervals in the late Paleocene/Early Eocene, the late Middle Eocene and the latest Eocene (Wolfe and Poore, 1982). Interpretation of flora suggests that during the Eocene, mean annual temperatures were high and the mean annual temperature range was small. The warm intervals resulted in the expansion of tropical vegetation to 50° latitude and broad-leaved evergreens to 70° latitude. Coniferous vegetation was restricted to the polar regions. The fossil remains of deciduous broad-leaved trees and coniferous forests at high latitudes on Axel Heiberg Island, with an accompanying diverse vertebrate fauna (West et al., 1981) suggests that the forest represents the Middle Eocene warm interval.

SUMMARY

Forests thrived on Axel Heiberg Island during the Eocene when the polar climate was much warmer. The forests grew on swampy floodplains between large river systems with frequently shifting courses so that the forests were flooded by sand, silt and mud. The sediment cover protected the leaf and wood material from decay and the vegetation is now preserved as spectacular mummified tree stumps, logs, and leaf litter layers.

The stumps are the remains of large forest trees that had trunks up to 1 m in diameter and grew to an estimated height of 20 to 30 m. Fossil leaves in the litter indicate that the trees in the forests included dawn redwood, swamp cypress, pine, fir and larch conifers, along with the angiosperms katsura and members of the walnut, birch and oak families.

Wide growth rings in the wood indicate that the trees grew fairly fast in a warm, moist environment. The productivity of the forests is estimated to have been about 1200 gm/m^2/y, comparable to that of living temperate deciduous forests. The density of the forests was also quite high, in the region of 325 to 484 trees a hectare. Despite the high latitude at which these forests grew they were adapted to growth in the polar light regime, probably by becoming dormant in the dark winter and growing rapidly during the summer. These polar forests are unique in that no living analogue exists on earth today, although if the increase in CO_2 levels in the atmosphere continues, forests may well develop in high latitudes once more.

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THE FOSSIL FORESTS OF THE BUCHANAN LAKE FORMATION (EARLY TERTIARY), AXEL HEIBERG ISLAND, CANADIAN ARCTIC ARCHIPELAGO: PRELIMINARY FLORISTICS AND PALEOCLIMATE

James F. Basinger[1]

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Abstract

The Eureka Sound Group of the Canadian Arctic Archipelago comprises a coal-bearing sequence with an early Tertiary fossil flora. Eocene sediments of the Buchanan Lake Formation are exposed on northeastern Axel Heiberg Island and contain an extraordinary fossil assemblage that includes in situ tree stumps. Stumps over one metre in diameter provide evidence of lush plant growth at high paleolatitudes.

Three distinct, facies-controlled macrofossil assemblages have been recognized within the sediments of the Buchanan Lake Formation; they apparently lack individual biostratigraphic significance and constitute a single regional flora. Dominant taxa include Taxodiaceae, Cercidiphyllaceae, Platanaceae, Betulaceae, Juglandaceae, Fagaceae, and Ginkgoaceae, representing a vegetation type common to high-latitude Paleocene and Eocene deposits and referred to as the "polar broad-leaved deciduous forest" or the "Arcto-Tertiary forest." However, the Buchanan Lake Formation assemblage also includes uncommon but diverse Pinaceae, indicating the presence of coniferous evergreen vegetation, and thus differs from other early Tertiary arctic assemblages, including those of the underlying Iceberg Bay Formation.

Floristics, features of vegetation, and wood anatomy indicate that the polar paleoclimate was mild and equable, with winter temperatures seldom below freezing, that there was an abundance of moisture, and that growth continued uninterrupted throughout the growing season.

Résumé

Dans l'Archipel arctique canadien, le groupe d'Eureka Sound englobe une séquence houillère caractérisée par une flore fossile du Tertiaire inférieur. Des sédiments éocènes de la formation de Buchanan Lake affleurent dans la partie nord-est de l'île Axel Heiberg, et contiennent un assemblage extraordinaire de fossiles, comprenant des souches d'arbres en place. Des couches de plus d'un mètre de diamètre témoignent de l'existence par le passé d'une végétation luxuriante à des paléolatitudes élevées.

La macroflore associée aux souches est dominée par des conifères et angiospermes à feuilles caduques, accompagnés de conifères rares ou localement abondants, à feuilles persistantes. Trois assemblages distincts, déterminés par des faciès, ont été identifiés. Apparemment, ils n'ont pas individuellement d'importance biostratigraphique, et constituent une seule flore régionale.

Le type de végétation représenté par les taxons dominants est commun aux dépôts paléocènes et éocènes des latitudes élevées. L'assemblage de la formation de Buchanan Lake comprend des *Pinaceae* peu communes mais diverses, qui indiquent l'existence d'une végétation de conifères à feuilles persistantes. Il diffère ainsi des assemblages d'autres localités arctiques, et de ceux de la formation d'Iceberg Bay sous-jacente. L'étude floristique, les caractères de la végétation, et l'anatomie du bois indiquent que le climat polaire se caractérisait par des températures hivernales rarement inférieures au point de congélation et par une saison de croissance ininterrompue, et aussi que l'humidité était abondante.

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[1] Department of Geological Sciences, University of Saskatchewan, Saskatoon, Saskatchewan S7N 0W0

INTRODUCTION

Early Tertiary coal-bearing strata and fossil floras were discovered throughout the High Arctic during nineteenth century polar exploration. Clearly these deposits represent periods in the Earth's past when the polar land masses experienced a much milder climate than at present. These northern fossil floras are remarkable for their uniformity over vast regions of the northern hemisphere. Similar assemblages have been recovered from Spitsbergen, Greenland, Ellesmere Island, Yukon, northeastern Asia, and the interior of North America from the Mackenzie Valley to the Great Plains. This vegetation of early Tertiary high latitudes is referred to as the Arcto-Tertiary forest (after Chaney, 1938, 1947, 1959), or the polar broad-leaved deciduous forest (Wolfe, 1985). Fossil materials from the Arcto-Tertiary forest were recovered from Ellesmere Island in the nineteenth century (Heer, 1878a, b; Brainard, 1886; Schei, 1903, 1904; Nathorst, 1915), and have been reported periodically in the twentieth century (McMillan, 1963; Tozer, 1963; Riediger and Bustin, 1987).

The locality on Axel Heiberg Island, known informally as the "fossil forest" or "Geodetic Hills fossil forest", was recognized during Geological Survey of Canada field operations in the summer of 1985 (79°55'N, 88°58'W; Fig. 1). Survey geologist B.D. Ricketts and helicopter pilot P. Tudge independently reported non-mineralized wood and in situ stumps. Earlier work of Balkwill and Bustin (1975), Hills and Bustin (1976), and Bustin (1982), in this and other areas, had revealed mummified wood, leaf litter mats composed primarily of *Metasequoia* and *Glyptostrobus*, and numerous seed cones of *Picea* and *Pinus*. Fieldwork for the present study was initiated in 1986 by J.F. Basinger, J.E. Francis, and E.E. McIver, with subsequent field seasons variously including these workers as well as B.A. LePage and S.E. Scheckler.

The fossil forests of Axel Heiberg Island represent a flora intermediate to those of the Arcto-Tertiary forests of the older, Iceberg Bay Formation, and the younger, Beaufort Formation. The Arcto-Tertiary vegetation was dominated by deciduous Taxodiaceae and angiosperms, including Cercidiphyllaceae, Platanaceae, Juglandaceae, and others, and bears more similarity to the modern vegetation of eastern Asia than that of North America. The macroflora of the Neogene Beaufort Formation of the Canadian Arctic reveals dominance by Pinaceae, with the occurrence of Betulaceae, Juglandaceae, and other angiosperms (Hills and Ogilvie, 1970; Roy and Hills, 1972; Hills et al., 1974; Matthews, 1987), and is fundamentally a boreal mixed evergreen coniferous/broad-leaved deciduous forest comparable to modern northern mixed forests of North America. The mix of typical (and dominant) Arcto-Tertiary elements with, for example, a rich diversity of Pinaceae, is distinctive among Canadian High Arctic macrofloras.

Materials recovered from the Axel Heiberg fossil forest site are exquisitely preserved and will contribute to an evaluation of taxonomic relationships of some elements of the high-latitude paleofloras, leading to a better understanding of their role in the evolution of members of modern temperate assemblages. In addition, host deposits representing various depositional environments are accessible, as a result of extensive exposure of unconsolidated sediments, so that both diversity and plant community structure can be studied. Data available should facilitate interpretation of the regional vegetational mosaic and the paleoenvironment.

This paper is a preliminary account of the fossil flora of the Buchanan Lake Formation of the Eureka Sound Group from collections made in the vicinity of the "Geodetic Hills fossil forest site". The flora and stump fields have been the subject of popular accounts (Basinger, 1986; Francis and McMillan, 1987), but have not yet been presented systematically. Recent work by Hickey et al. (1983), McIver and Basinger (1989), and Lepage and Basinger (in press, and *this volume*) represents ongoing investigation of the floras of this group. A thorough, systematic account of the flora is beyond the scope of the present paper, and is, in part, the subject of the doctoral dissertation of B. A. LePage. An overview of the flora is presented here and some of the more significant taxa are illustrated.

For a detailed sedimentological profile of the site, the reader is referred to the work of Ricketts (1986, 1987, *this volume*) and Ricketts and McIntyre (1986), and references therein.

MATERIALS AND METHODS

Geology

Investigations by Irving and Wynne (*this volume*) show that Axel Heiberg Island has moved north at most a few degrees of latitude since Eocene time, supporting McKenna's (1980) earlier estimate. Thus, the Buchanan Lake Formation sediments of Axel Heiberg Island were deposited in a region located well above the Arctic Circle. The forests growing on these sediments would have been subjected to as much as three months of continuous summer daylight and three months of winter darkness annually.

The Eureka Sound Formation was established by Troelson (1950), Tozer (1963), and Souther (1963) to include deposits of sandstone, shale, and lignite that are found widely scattered in the Canadian Arctic Archipelago. Sediments of the Eureka Sound Formation were distinguished from those of the younger Beaufort Formation on the basis of the style of preservation of fossil woods, usually permineralized in the former, and unaltered and uncompressed in the latter, and by the flat-lying and undeformed (post-orogenic) nature of the Beaufort deposits. At many localities Eureka Sound sediments were folded during the Eurekan Orogeny.

The Eureka Sound Formation was elevated to group status independently by Ricketts (1986) and by Miall (1986). Both authors included syn- and post-orogenic deposits of northeastern Axel Heiberg Island within the Eureka Sound Group, and named these deposits the Buchanan Lake Formation (Ricketts, 1986), and the Boulder Hills Formation (Miall, 1986). Ricketts examined the sediments in the vicinity of the Axel Heiberg fossil forests during the present study, and Ricketts' stratigraphic nomenclature is used in this paper. The four formations named by Ricketts, in ascending order, are: Expedition Formation, Strand Bay Formation, Iceberg Bay Formation, and Buchanan Lake Formation.

The Buchanan Lake Formation was deposited in a northwest trending depression that formed at the foot of the rising Princess Margaret Range of mountains during the final stages of the Eurekan Orogeny (Tozer, 1960; Bustin, 1982; Ricketts, 1986, 1987, *this volume*; Ricketts and McIntyre, 1986). Fossil plant remains are contained within a unit of fine grained channel and floodplain sediments lying conformably on thick conglomerates (Ricketts and McIntyre, 1986; Ricketts, *this volume*). For the purposes of this paper, the channel and floodplain sediments are divided into three units: a lower, finer grained unit, about 50 m thick, containing some leaf-litter mats and tree stumps; a middle, heterogeneous unit with a moderate proportion of crossbedded sands, also about 50 m thick, which contains most of the litter mats and stumps; and an upper, coarser unit, nearly 100 m thick, which is composed mainly of sand and includes many "channel sands assemblages" but few litter mats. These divisions are made for convenience of paleobotanical description and are not meant to imply stratigraphic significance. The increase in the

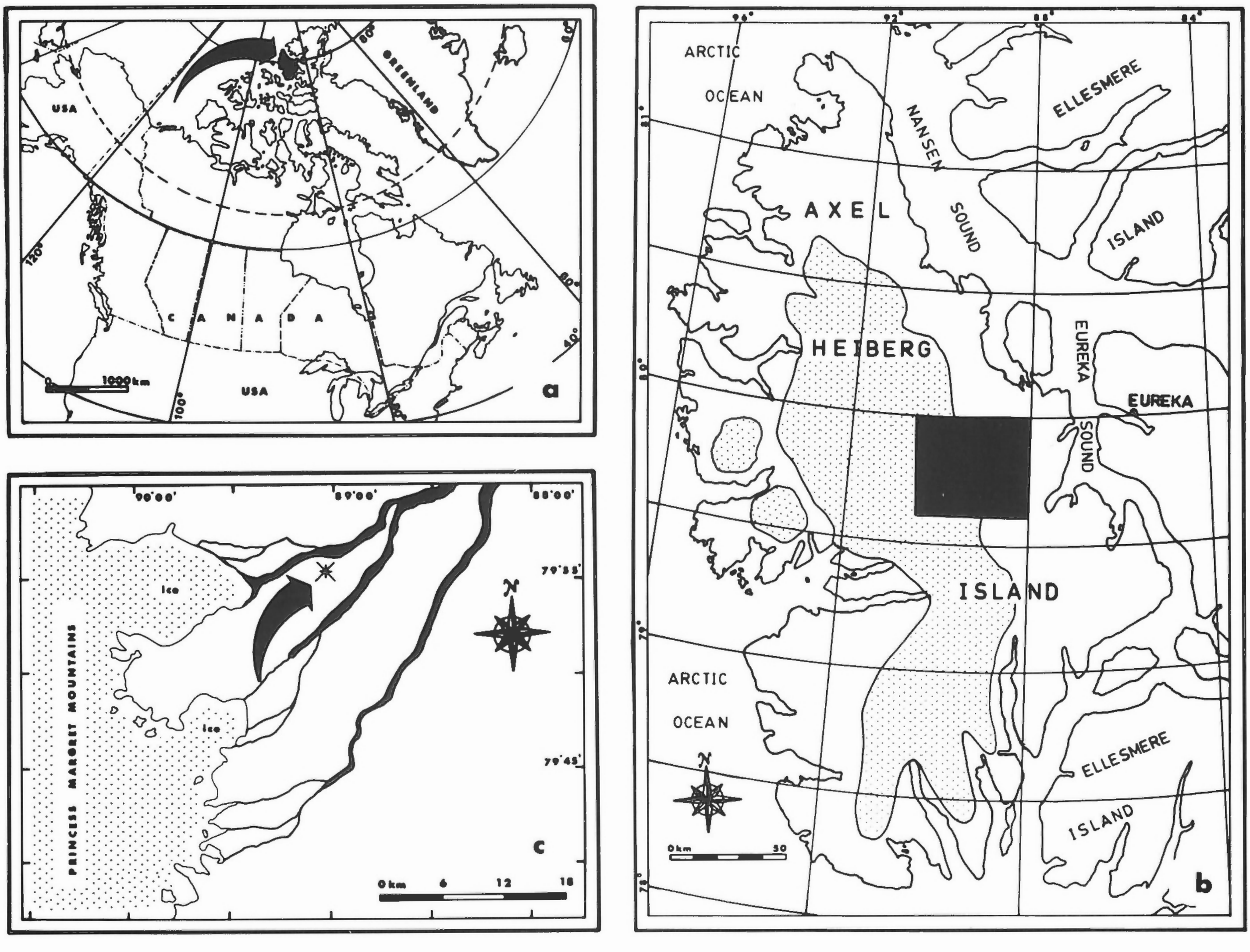

Figure 1. *a, map of Canada showing the location of Axel Heiberg Island; b, detailed map of Axel Heiberg Island showing location of fossil forest area; c, detailed area map showing fossil forest site. Stipple represents ice.*

proportion of channel sands and a decrease in floodplain silts and plant matter upward coincides with upwardly decreasing diversity and abundance of plant fossils.

The lower, fine grained unit is not well exposed at the fossil forest site, but is accessible at a stream bank exposure approximately four kilometres north-northwest, on a tributary of the major river to the north. Active erosion by glacial meltwater maintains fresh exposures of poorly consolidated siltstone horizons at this locality. This type of sediment, which rapidly breaks down when exposed to weathering, is not available for sampling at other localities, and contains an assemblage of deciduous angiosperm fossils not found elsewhere.

The fine state of preservation of the plant remains is in part due to shallow burial. Whether there is a significant difference in the degree of maturation of organic matter from the bottom to the top of the fine grained unit is not known. Leaf litter appears to be more coaly in the lower beds, but well preserved mats may be found throughout the section.

A coarse, boulder gravel deposit, containing a bed of fossil mosses that indicate a late Neogene age (C. LaFarge-England, pers. comm.), unconformably overlies the Buchanan Lake Formation sediments. Incision and formation of the modern landscape has left these gravels capping only the highest hills in the area. Where these deposits have been removed an obscuring boulder "drape" may remain. Although the maximum thickness of these gravels cannot be determined, it is unlikely that they exceeded 100 to 200 m, if maximum modern relief in the area can be viewed as limiting. Although maximum depth of burial was not great, the lowermost sediments would have been buried to twice the depth of the higher forest levels (200 to 400 m vs. 100 to 200 m), which may have contributed to greater compaction of the basal deposits.

Age

The age of the Buchanan Lake Formation, now considered Eocene by McIntyre (Ricketts and McIntyre, 1986; McIntyre, *this volume*), has been the subject of some dispute. Mummified wood and well preserved seed cones of *Picea* from exposures in the area of the fossil forest were interpreted by Hills and Bustin (1976) and by Bustin (1982) as indicative of Beaufort beds and a Miocene age. Although the Buchanan Lake Formation wood is mummified or unaltered, it is more compressed and appears far less "fresh" than is typical for wood recovered from the Neogene sediments of the Beaufort Formation. The assignment of an age from a single taxon must be considered uncertain, or tentative.

Ricketts and McIntyre (Ricketts, 1986, 1987; Ricketts and McIntyre, 1986) reevaluated the relationships of sediments associated with the Stoltz Thrust Zone of northeastern Axel Heiberg Island, and assigned sediments, including those at the fossil forest site, to the newly recognized Buchanan Lake Formation. They concluded that the Buchanan Lake Formation was deposited during and after activation of thrusting associated with the uplift of Princess Margaret Arch, during the final stages of the Eurekan Orogeny. Based on structural and palynological evidence, they estimated that this uplift occurred during the Middle or Late Eocene.

Ricketts and McIntyre (Ricketts, 1986, 1987, 1988; Ricketts and McIntyre, 1986; (McIntyre, *this volume)* assigned a Middle Eocene age to that part of the Buchanan Lake Formation that includes the fossil forests, although the microflora does not preclude a Late Eocene age (McIntyre, *this volume*). According to Miall (1985, 1986, 1988), however, there is no evidence precluding an age as young as Early Oligocene.

Norris (1982, 1986) noted significant and rapid palynofloral change, including extinction of many taxa characteristic of the Eocene, associated with the Eocene/Oligocene boundary in the Mackenzie Delta region of the Northwest Territories. This palynofloral change in the Canadian Arctic should preclude the possibility of a post-Eocene age for the fossil forests. Definition of the Eocene/Oligocene boundary and correlation of Norris' floristic break are beyond the scope of this study, but it would seem that this break is related in some way to the so-called terminal Eocene climatic event (see Pomerol and Premoli-Silva, 1986). Oligocene macrofloras are unknown from the Canadian Arctic. This break is apparent in the contrast between the Neogene Beaufort floras and the Paleogene, Arcto-Tertiary, Eureka Sound floras, as noted above.

A search was made for vertebrate remains at the fossil forest section by M.R. Dawson, M.C. McKenna, H. Hutchison, and C. Howard in 1987, but none were found. Bone fragments recovered by Dawson's party from deformed sediments of the Iceberg Bay Formation (Eureka Sound Group; stratigraphically beneath the Buchanan Lake sediments) outcropping near Mokka Fiord, about 50 km southeast of the forest site, are comparable to vertebrate remains from Iceberg Bay deposits of Ellesmere Island (M.C. McKenna, pers. comm., 1987). These faunas are indicative of an early or early Middle Eocene age (West et al., 1977; Estes and Hutchison, 1980; McKenna, 1980).

No igneous rocks have been found in the area and no absolute ages are available.

Collection, curation, and preparation

The fossil forest site is exposed on the flanks and top of a low, teardrop-shaped hill whose long-axis trends roughly northeast. Extraordinary eolian erosion, a result of local geography and the orientation of the hill relative to the prevailing northeast winds, constantly strips sediment from the hill and prevents the formation of soil even though the top of the hill is quite flat. The tree stumps are more resistant to wind ablation than the enclosing sediments, and so they stand out in relief as sediment is scoured from around them (Plate 1, fig. 1). The leaf-litter mats also resist wind ablation to form benches on the flanks of the hill and wide areas of exposure on the brow of the hill. Disturbance of these litter mats may make them vulnerable to erosion. Elsewhere, active erosion of steep valley walls results in fair exposure, but access to fossiliferous beds is limited by steepness of slope and shallow permafrost (typically 20 to 30 cm beneath the surface). Except for the exposures described, frost action thoroughly disturbs the active layer and obscures parent material. The fossil forest site proper exposes only the middle sedimentary unit, which contains the majority of the leaf-litter mats. On the adjacent, much higher, hill there are exposures of the upper part of the middle unit and all of the upper, sandy unit.

Exploration of the region has so far failed to reveal any other areas comparable in productivity to the fossil forest site. This site may prove to be unique within the Buchanan Lake Formation both in terms of the concentration of fossil beds and in the extent of their exposure.

The woody remains were found to be water saturated and vulnerable to differential shrinking and cracking when removed from the sediment. Specimens exposed at the surface for some years have dried naturally, however, and appear more or less immune to further damage. Such specimens curated at the University of Saskatchewan have received no special treatment and show no signs of deterioration.

Leaf litter and conifer cone materials were found to be exceedingly fragile when dry. Leaf-litter mats were examined for specimens of cones, leaves, fruits and seeds. In all cases where litter mats were found to be productive, bulk samples were also removed. Bulk samples were either packed into cotton sacks or wrapped tightly in newspaper. Bulk samples were air dried in the laboratory in preparation for picking.

Fragile cones, fruits, and other small specimens presented special transportation problems. Even the wrapping of specimens in fine tissue caused crushing, so this method was abandoned. It was found that specimens were best protected when packed in a rigid container filled with fine, dry sand. The sand surrounds the specimens and fills cavities, protecting the specimens against movement and concussion, and prevents crushing. The sand is easily removed on unpacking. The most practical containers were found to be empty food cans, as these were a convenient size and easily closed and sealed with tape.

Air drying of leaf and cone material seemed to have little effect on specimens, other than loss of flexibility and increased fragility. Sediment clinging to specimens could be removed with an overnight treatment of full-strength (52%) hydrofluoric acid (HF), which greatly improved the appearance of specimens for examination and photography.

Woods were embedded in glycol methacrylate prior to sectioning on a steel microtome.

All specimens collected during this study and illustrated in this report are housed in the University of Saskatchewan Paleobotanical Collection (USPC), the Department of Geological Sciences, University of Saskatchewan, Saskatoon, Saskatchewan, Canada S7N 0W0.

COMPOSITION OF THE FLORA

Most fossil tree stumps occur in a small number of discrete beds, each "forest" of stumps apparently representing a distinct episode of forest development and simultaneous preservation (Francis and McMillan, 1987; Francis, *this volume*). Some fossil forests are rooted in leaf mats; others are preserved with root systems radiating into more oxidized paleosols. Whether this has paleoecological significance is not known. Logs, some up to ten metres in length and nearly one metre in diameter, are found associated with stumps and leaf mats as well as in isolation.

Remains of fossil plants, other than stumps and logs, occur in three main depositional settings: leaf litter, fluvial sands, and siltstone. The leaf-litter floras are autochthonous accumulations of debris that represent the plants of the swamps or peat-forming environments (Plate 2, fig. 1). The floras of the channel sands include transported material, mostly wood, cones, and seeds, carried by water for some distance before burial. The siltstone floras are found in fine grained floodplain deposits and represent locally derived material, probably from plants that grew on the floodplain and were transported short distances before being buried. The floristic composition of these deposits is summarized in Table 1.

The leaf-litter floras

Vegetative and reproductive remains are clearly recognizable in most organic beds and in some cases are exquisitely well preserved. Those mats with a minor component of silt coating the plant remains tend to yield the best material, although the presence of silt cannot be correlated with any differences in floristic composition. The presence of silt may lessen the effects of peatification, thereby improving the quality of preservation and facilitating the isolation and removal of fossil specimens.

The litter mats range from several centimetres to nearly one metre in thickness, and many beds can be traced for more than one kilometre. Fine preservation is not confined to the upper part of a mat but may occur throughout the mat, indicating a water table that was at or above the surface of the ground for most of the year, so that biodegradation was limited. Accumulation of thick and laterally extensive mats of plant remains, embedded roots, in situ stumps, and the common absence of inorganic sediment, all indicate an autochthonous origin for the leaf mats. The leaf mats are interpreted as representing extensive lowland swamps.

TABLE 1

Distribution of taxa found in the Buchanan Lake Formation according to facies-controlled assemblage type

Taxon	Abundant	Locally common[1]	Rare[2]
Leaf litter floras			
Metasequoia sp	•		
Glyptostrobus sp.	•		
Alnus sp.		•	
Betula sp.			•
"*Cercidiphyllum*" sp.			•
Carya sp.			•
unidentified angiosperm leaf types ("A", "B", "C")			•
unidentified angiosperm fruits and seeds (several taxa)			•
Larix sp.			•
Picea sp. (small cones)			•
Pinus sp.			•
?*Chamaecyparis* sp.			•
?*Tsuga* sp.			•
?*Abies* sp.			•
?*Pseudolarix* sp.			•
undescribed Pinaceae "A"			•
undescribed Pinaceae "B"			•
undescribed Taxodiaceae "A"			•
Osmunda sp.			•
unidentified fern "A"			•
Channel sands floras			
Picea sp. (large cones)	•		
Picea sp. (small cones)	•		
Metasequoia sp.		•	
Pinus sp.			•
Carya sp.			•
unidentified angiosperm fruits and seeds (few taxa)			•
Siltstone floras			
Many of the following taxa are site specific; relative abundance has not yet been determined:			
Betulaceae			
"*Cercidiphyllum*" sp.			
Platanus sp.			
Quercus sp.			
unidentified angiosperm leaf types (several)			
Metasequoia sp.			
Glyptostrobus sp.			
Ginkgo sp.			
unidentified ferns (2 taxa)			

[1] typically common or abundant at only a few localities.

[2] present as a minor component at a few localities.

Most of the remains found in the leaf litters are leafy shoots of the deciduous taxodiaceous conifers, *Metasequoia* (Plate 2, figs. 2,5) and *Glyptostrobus* (Plate 2, figs. 3,4). Virtually all leaf-litter layers include *Metasequoia* (dawn redwood), and most are dominated by it. *Glyptostrobus* (swamp cypress) commonly occurs as a minor component, rarely as a dominant one.

Evidence of other plants in the Axel Heiberg fossil swamp communities is rare. Understory or herbaceous plants include royal fern (*Osmunda*) and perhaps the few taxa preserved only as seeds, as yet unidentified. Rarely, the peat-forming communities were dominated by alder (*Alnus*) of the Betulaceae (Plate 4, fig. 2); the pollen flora of such mats is dominated also by *Alnus* (D.J. McIntyre, pers. comm.). Other rare components of these communities include a 3-needle pine (*Pinus*) (Plate 3, fig. 5), spruce (*Picea*, small cones), larch (*Larix*) (Plate 3, fig. 4), cedar (?*Chamaecyparis*) (Plate 3, fig. 6), (?)fir (?*Abies*), (?)golden larch (?*Pseudolarix*), (?)hemlock (?*Tsuga*), birch (*Betula*) (Plate 4, fig. 5), katsura ("*Cercidiphyllum*" fruits) (Plate 4, fig. 3), and extinct genera of Pinaceae and Taxodiaceae (Plate 3, fig. 7). Most represent undescribed taxa, and confirmation of generic identification of some taxa awaits further study.

Most of the taxa described as rare have been found in significant numbers in only a few litter layers, notably level "N" of Francis (*this volume*), within the lower to middle exposures at the site, representing the greatest concentration of litter beds. It is possible that these rare taxa represent transported materials. However, these taxa are commonly concentrated in "pockets", a few centimeters thick and only a few square metres in area, within individual litter mats. This, together with the observation that most of these taxa have living wetland relatives, suggests that their remains are autochthonous and that they periodically flourished in a type of peat-forming community.

It is important to note that, considering the scarcity of remains other than those of the taxodiaceous conifers within the leaf litter, superficial collection may be expected to result in a serious underestimation of diversity.

The fluvial sands floras

Remains preserved in channel, point bar, and related deposits are more abundant in the upper half of the section (the upper middle unit and upper unit) where sediments are coarser and fluvial sands are more common. The fluvial sands are mostly barren or contain rare organic debris, but, in places, these sands are rich in woody detritus. Most of the debris consists of decorticated axes up to a few centimetres in diameter. Resistant reproductive structures such as conifer cones and angiosperm nuts are also preserved, but remains as delicate as leaves have not been found.

The most common reproductive organs found in these deposits are cones of spruce (*Picea*). Two distinct size classes of *Picea* cones have been recovered; the larger are long and slender and up to 12 cm in length (Plate 3, fig. 1); the smaller cones are up to 5 cm in length (Plate 3, fig. 2). Reliable macrofossil remains of *Picea* have not been reported previously from pre-Oligocene deposits (Miller, 1970, 1977, 1989; Crabtree, 1983). Miller has shown, however, that the genus existed in the Early Oligocene and noted that "winged seeds, foliage, pollen grains, and occasional seed cones are well known from the Late Eocene onward" (Miller, 1989, p. 747). Miller went on to suggest that "*Picea* probably evolved during the latest Cretaceous or Paleocene" (op. cit., p. 753). The spotty and unreliable pre-Oligocene record could be accounted for if *Picea* had evolved in cool regions of high altitude and latitude. The conditions in regions such as these are conditions to which many species of *Picea* are presently adapted.

Cones of *Pinus*, similar to those that occur in the leaf litter, are uncommon at most sites (Plate 3, fig. 3). Hickory nuts (*Carya*) (Plate 4, fig. 7) are very rare in these deposits except at one site, no more than one square metre in area, where several hundred nuts were recovered. The host sediment is a lens of silt in a metre-thick heterogeneous bed comprising a number of small lenses of silt and sand, some rich in woody debris. Although the bed is well exposed, an intensive search produced only a few additional specimens. Apparently, taphonomic processes had concentrated these nuts within a small pocket. These tough fruits could have been transported many kilometres, although their occurrence in such a dense concentration would argue for a more local source and limited transportation.

Spruce cones from Buchanan Lake Formation sediments were illustrated by Hills and Bustin (1976) and Bustin (1982) and identified as *Picea banksii* Hills and Ogilvie, although Hills and Bustin (op. cit.) noted morphological dissimilarities. This identification, and its implications for the comparison of the floras from the Buchanan Lake and Beaufort formations, must be considered tentative until detailed anatomical and morphological studies are completed. Hills and Bustin (op. cit.) also reported, but did not illustrate, single specimens of *Metasequoia*, *Alnus*, and *Carya*. Bustin (op. cit.) reported the presence of *Metasequoia*, *Larix*, and *Alnus*.

The siltstone floras

Plant-bearing beds are well exposed in the banks of a stream about 4 km north-northwest of the fossil forest site. Glacial meltwater undercuts the banks, which then spall and collapse. The fossiliferous silts are poorly consolidated and are accessible only here, where fresh and commonly still frozen blocks are available. These beds immediately overlie massive conglomerates of the Buchanan Lake Formation and are considered to be at the same stratigraphic level as the lower beds (lower unit) at the fossil forest site (B.D. Ricketts, pers. comm.; D.J. McIntyre, pers. comm.).

The flora of the siltstones consists almost exclusively of broad-leaved deciduous plants. Oak (*Quercus*) (Plate 4, fig. 1), Betulaceae, sycamore (*Platanus*), maidenhair (*Ginkgo*) (Plate 4, fig. 6), *Metasequoia*, *Glyptostrobus*, "*Cercidiphyllum*", *Osmunda*, and several unidentified angiosperms have been recovered. The quality of preservation suggests limited transportation, and the remains probably represent floras that inhabited the better-drained soils of the floodplain. Taxa occurring in the siltstone facies show considerable site-specificity, and relative abundance is not easily determined on the basis of the available data. Leaf litter from this site is dominated by *Metasequoia*, and is comparable to that of the fossil forest site. The apparent absence of *Picea*, *Pinus*, or other Pinaceae present in other facies is noteworthy.

DISCUSSION

Age of the fossil flora

Before the flora of the fossil forest site can be interpreted, it must be established that the fossil remains are not disparate in age. There is no sedimentological, palynological, or other evidence of a substantial hiatus within the host strata (Ricketts and McIntyre, 1986; McIntyre, *this volume*; Ricketts, *this volume*). The macroflora of the deposit includes assemblages from three distinct facies types. Because the sediments become sandier toward the top of the section, the vertical distribution of certain facies and their associated plant remains is uneven. Therefore, some taxa of the siltstone floras have not been found above the lower, fine grained beds. Spruce cones, on the other hand, occur nearly exclusively in the channel sands, and are most abundant in the upper, coarser beds.

The following observations may be made:

1. Leaf-litter mats are found throughout the section (except in the uppermost beds of the upper unit) and are composed primarily of *Metasequoia*. With the exception of *Glyptostrobus*, other taxa occurring in the litter mats typically have been found at only one or at most a few localities. Many of the rare taxa were found in a portion of a single mat (the "N" layer of Francis, *this volume*), in the middle of the greatest concentration of litter mats. Adjacent litter mats (within two metres above and below), as well as part of "N" layer itself, are taxodiaceous. No discernible pattern is apparent, other than simply that

diversity is greatest where mats are most abundant and best exposed. The "N" layer is paleoecologically distinct, but at this time cannot be correlated with regional climatic changes.

2. Spruce cones occur in sandy deposits throughout the middle and upper units. Although two size classes of cones are recognized, the taxonomic affinities of cones throughout the section have not yet been determined, in part due to the highly variable quality of preservation.
3. The siltstone floras of the lower unit include species of *Metasequoia*, *Glyptostrobus*, "*Cercidiphyllum*", *Osmunda*, and Betulaceae, in common with the litter mats of the forest site. Siltstone floras are also associated with taxodiaceous litter mats. Ironstone nodules, which are widely scattered and uncommon at the forest site proper, contain *Platanus*, Betulaceae (Plate 4, fig. 4), and *Ginkgo*, further illustrating the floristic relationships between the lower and middle units.

It should be noted that the macroflora in the upper 50 m of the sandy, upper unit is poorly known, because of poor recovery of fossils and the upward loss of all but the channel sand assemblages. Taxa found at this level appear conspecific with those of lower levels. Variability in cone size has been noted for *Metasequoia* (L.V. Hills, pers. comm.) and *Picea*. However, in the current study, cone size has been found to be stratigraphically inconsistent, and all cones of *Metasequoia* are within the intraspecific limits for *M. occidentalis* (Newberry) Chaney, as defined by Chaney (1951) and Christophel (1976).

Macrofossil evidence is insufficient to permit the recognition of floristic zonation within the section. Inconsistent recovery of taxa appears to be controlled primarily by a combination of original sedimentological and recent erosional processes, which has reduced the availability of facies-controlled assemblages. The macroflora, with the majority of taxa occurring within a 30 m thick interval in the middle unit, is here considered collectively to be a single flora. The flora is interpreted as having been influenced by the geographic evolution of the basin, but does not indicate major floristic or climatic changes during deposition. This is compatible with the palynological interpretations of McIntyre (Ricketts and McIntyre, 1986; McIntyre, *this volume*).

The flora of the Paleocene-Eocene Iceberg Bay Formation of Axel Heiberg and Ellesmere islands, as listed by Hickey et al. (1983) and as found by the author, shows similarity at the generic level to the Buchanan Lake Formation flora. The dominant vegetation type in both formations is Arcto-Tertiary in aspect, but the presence of taxa such as diverse Pinaceae and lobed oaks distinguishes the Buchanan Lake flora from known Iceberg Bay floras. Neogene forests of the Arctic Islands, as typified by Beaufort assemblages (Hills and Ogilvie, 1970; Roy and Hills, 1972; Hills et al., 1974; Matthews, 1987), were dominated by Pinaceae and were no longer of the Arcto-Tertiary type.

The Buchanan Lake Formation flora shows evidence of the transition of high-latitude vegetation from the Arcto-Tertiary (Iceberg Bay) type to the Boreal (Beaufort) type. A pre-Oligocene age for the Buchanan Lake flora has been determined on the basis of palynology by McIntyre (Ricketts and McIntyre, 1986; McIntyre, *this volume*). A post-Eocene age may also be inconsistent with the vegetational and climatic changes elsewhere associated with the "Terminal Eocene Event" (Wolfe, 1978, 1985, 1987; Norris, 1982, 1986; see also Pomerol and Premoli-Silva, 1986, and papers therein). Although McIntyre stated that the palynoflora is most likely of Middle Eocene age and that there is no evidence that even the uppermost beds of the Buchanan Lake Formation are as young as Late Eocene, he did indicate that a Late Eocene age is not necessarily precluded (Ricketts and McIntyre, 1986; McIntyre, *this volume*). Floristically and vegetationally, the Buchanan Lake macroflora is compatible with a Middle or Late Eocene age. The contrast between the Buchanan Lake flora and the Iceberg Bay flora, however, supports a post-Iceberg Bay, and possibly a Late Eocene, age. The possibility that part or all of the Buchanan Lake Formation may be Oligocene in age has been suggested (Miall, 1985, 1986, 1988; L.V. Hills, pers. comm., 1988; A.R. Sweet, pers. comm., 1990). Although the macroflora may not preclude such an age, there is as yet no compelling evidence for a post-Eocene age.

The Buchanan Lake Formation flora can be used to illustrate a hazard in macrofloral age determination. The flora of the channel sands is dominated by *Picea* and most closely resembles the Neogene Beaufort assemblages. The vast majority of leaf-litter mats, however, are indistinguishable floristically from similar mats of Paleocene and Eocene age. Because floral diversity is strongly influenced by facies, limited sample size or access to only a single facies type may lead to erroneous concepts of the regional paleoflora and may contribute to misinterpretation of age.

Fossil wood and the preservation of organic remains

Fossil stumps and logs are commonly concentrated in relatively thin beds and are not randomly scattered throughout the sedimentary column. Many such beds, here called fossil forests, have been recognized at the fossil forest site; they contain a few to numerous stumps and associated wood enclosed in massive bedded silt and sand (Francis and McMillan, 1987; Francis, *this volume*). The stumps are variously rooted in organic paleosols (leaf litter) and in silty, poorly developed paleosols (Foscolos and McMillan, *this volume*; Tarnocai et al., *this volume*). The trees evidently grew in an environment that is interpreted as an extensive floodbasin, with the enclosing sediments deposited as crevasse splays (Ricketts, *this volume*). The uniformity of preservation of stumps, the regularity of their spacing, and the massiveness of the enclosing sediments are consistent with rapid burial of each forest layer.

Analysis of the mummified wood reveals extremely little chemical alteration or biological degradation (Obst et al., *this volume*), indicating that the preserved stumps were at all times below the water table. The condition of many leaf-litter mats associated with the stumps also indicates that decay of material on the forest floor was commonly inhibited, again suggesting that the water table was frequently at or near the sediment surface. Sediment that killed a stand of trees by suffocation could cause a rise in water table by the raising of the land surface and related capillary action. Under the right conditions this could preserve the stumps. Dead trunks protruding above the sediment would soon decay to a level roughly corresponding to the water table. This would explain the consistent height of stumps within a single bed and account for their not protruding into succeeding beds.

The lack of significant ash in the mummified wood (Obst et al., *this volume*) further constrains interpretation of the conditions of burial. Wood has a strong affinity for dissolved minerals and commonly is permineralized if infiltrated by groundwater (Leo and Barghoorn, 1976). Minimal permineralization of fossil forest wood may be explained by entombment within fine grained silts that limited groundwater movement. Permineralized woods are found at the top of the section at the fossil forest site, however, where they are associated with coarser grained sediments.

Anatomical preservation of the fossil forest woods is extremely variable, due primarily to the degree of deformation of the tissues during compression. Logs lying horizontally are commonly compressed by a factor of 4:1 or more. For reasons that are not clear, some woods have suffered severe compression, while others at more or less the same stratigraphic level are scarcely deformed (see Plate 1, fig. 2). Preservation is in some cases adequate for identification at the family level, but the details of cell walls required for more critical identification are difficult to assess and features of annual rings are not clearly recognizable.

Tree stumps, as expected, have resisted crushing more effectively due to their vertical orientation. Nevertheless, examination of most stumps reveals considerable loss of diameter, in some cases estimated to be as much as 50 per cent, through radial crushing of tissue. This crushing is seen in the collapse of tracheids and is commonly expressed in the typical chevron appearance of the wood, resulting from diagonal distortion as cell walls collapsed. Horizontal compression of the stumps indicates instability or perhaps plasticity of the sediments during compaction. Although most stumps and logs are not well preserved anatomically, a few specimens show fine preservation.

Wood identification and forest composition

The fossil stumps are autochthonous and form the fossil forests proper. Isolated logs and wood, particularly those in the channel sands, are most likely allochthonous and may be more representative of the regional vegetation. On the basis of variations in the shapes of stumps and root habits, Francis (*this volume*) believes that more than one taxon is represented in the forests. Obst et al. (*this volume*) and Young (*this volume*) have determined that several taxa of the Pinaceae and Taxodiaceae may be present among the woods, but it is important to establish which of their specimens represent fossil stumps. In the present study, only one of the isolated logs and none of the stumps possess both the vertical and horizontal resin canals that conform with the resinous Pinaceae. Woods associated with channel sands were not examined in the present study, but may be predicted to be rich in *Picea* and *Pinus*, judging by the presence of cones of these taxa in the associated debris. Remains of angiospermous woods have not been found.

Distinction among the woods of most Taxodiaceae, Cupressaceae, and some Pinaceae is difficult and depends upon such features as microscopic details of ray cell and wood parenchyma cell walls – if it can be assumed that such features are consistent within taxa (Torrey, 1923; Greguss, 1955; Basinger, 1981). Differentiation of woods belonging to certain genera, such as *Sequoia*, *Metasequoia*, and *Sequoiadendron*, may not be possible (Greguss, 1955; Schwarz and Weide, 1962; Basinger, 1981). This considered, it may be impossible to demonstrate either the generic affinities of the stumps, or whether they belong to one or more taxa.

Most of the leaf-litter floras are dominated by *Metasequoia*. *Glyptostrobus* is of secondary importance. While some litter shows a mixing of the two, these taxa commonly appear to have formed monotypic stands. The lowland swamp forests that formed the litter mats only rarely included other taxa (see Table 1). The close association of fossil forests and litter mats, with some stumps rooted within the mats, suggests that the stumps, too, may represent monotypic stands of taxodiaceous conifers. As the dominant plants of the wet lowlands, the taxodiaceous conifers would have contributed most material to the leaf mats and should comprise the bulk of the material buried by crevasse splay deposition. If it can be demonstrated that the stumps are not taxodiaceous, then there would be an implied paleoecological difference between the litter-mat swamp forests and the forests preserved as stumps (the fossil forests proper).

The monotypic forests forming the leaf litters may be considered analogous to modern taxodiaceous swamps, such as the bald cypress (*Taxodium*) swamps of southeastern North America. Saturated soils and frequent flooding in such habitats discourage development of an understory, which may in part explain the absence of herbaceous remains in the fossil leaf litters (rare *Osmunda* being an exception).

Paleoclimatic implications

In four well preserved whole or fragmentary stumps, ten or more largely undeformed annual increments of growth are identifiable. Annual increments of growth, or annual rings, of these fossil woods are up to 7 mm in width, and ring width tends to be uniform within a specimen. Plate 1, figure 2, illustrates wood from an exceptionally well preserved specimen and reveals a single, broad ring from a sequence of more than thirty rings. This specimen shows straight ring boundaries throughout its approximately half-metre length, and little tangential curvature of rings, confirming that the specimen is of mature trunk wood.

Tracheids were found to vary little in radial width from initiation of growth to within a few cells of the end of the ring, where lumen radial diameter is abruptly reduced to near zero. Using criteria of lumen size and shape (see discussion by Creber and Chaloner, 1984), only those few cells at the margin of the ring can be recognized as late wood. Creber and Chaloner noted the difficulty inherent in recognition of the early/late wood transition, and the many quite different criteria on which recognition of this transition is based. Severe crushing in most of the Buchanan Lake Formation woods exacerbates the problem. Even in the best preserved woods, microscopic examination reveals that cell walls are no longer straight, but meander, as if the wood at some time during burial lost its strength and the cell walls "sagged".

The first-formed cells are the weakest cells of the annual ring and are the first to collapse during crushing. This crushed zone of early cells tends to accentuate the ring boundary because crushed early wood cells may macroscopically mimic late wood. The degree of crushing of early wood is highly variable, from the minimal distortion seen in Plate 1, figure 2, to the collapse of nearly all cells. Within a single specimen, crushing may vary from ring to ring. The boundary between crushed and intact cells is related to the relative strength of cell walls and to the force of compression. This boundary should not be interpreted as an early/late wood transition.

The growth rings in wood from the Axel Heiberg fossil forest are relatively large when compared with other fossil woods (see Creber and Chaloner, 1984), supporting the observation by Parrish and Spicer (1988) that no latitudinal gradient in ring width existed at that time. This contrasts strongly with the climate-related latitudinal gradient in ring width as seen in the modern world (Creber and Chaloner, 1984, 1985).

Wide annual rings showing little variation in tracheid size (no signs of stress, such as false rings) and with very little late wood have been reported in fossil woods found in the high latitudes of North America and Antarctica (Jefferson, 1982; Creber and Chaloner, 1984; Francis, 1986, 1988; Parrish and Spicer, 1988). These authors have interpreted this feature as an indication that growth proceeded unhindered throughout the growing season and that growth was terminated abruptly by low light conditions – climatic conditions were otherwise not limiting when growth ceased. When interpreting tree growth on the basis of wood recovered from the fossil forests proper, it must be kept in mind that the trees probably represent only one microhabitat, the lowland swamps. It is unlikely, even in years of reduced rainfall, that these trees would have experienced much water stress. While ring width and the uniformity of wood development have been used to develop an interpretation of the climatic parameters for the growing season, the near-absence of late wood among high-latitude fossil woods has not been fully explained.

Relative abundance of late wood in annual rings has been attributed to such factors as tree age and ecological changes during the lifetime of the tree, temperature changes during the growing season, distribution of precipitation during the growing season, or a genetic predisposition for little or excessive late wood production (Tomlinson and Craighead, 1972; Creber, 1977; Creber and Chaloner, 1984; Francis, 1986). It is important to consider wood development as two separate processes, however: initiation of late wood production, and the actual production of late wood.

Growth ring formation (e.g., Plate 1, fig. 3) is a cambial response to hormonal changes occurring within the plant (Larson, 1962; Zimmermann and Brown, 1971). The hormonal changes that affect cambial activity are those that influence all parts of the tree and include the actions of auxins, giberellins, and other growth promoters and inhibitors (Zimmermann and Brown, 1971). While relative abundance of these compounds may be influenced by water or temperature stress (forming false rings, for example), the seasonal and systemic morphological, anatomical, and biochemical changes, of which late wood is but one component, are typically under photoperiodic control. In temperate regimes, these changes are part of the hardening process that prepares a plant for dormancy.

Short daylength is recognized as a primary trigger in the series of events leading up to winter hardiness (Larson, 1960; Irving and Lanphear, 1967; Weiser, 1970; Li and Palta, 1978; Fuchigami et al., 1982; Sakai and Larcher, 1987). Low temperatures and biological clocks can also induce cold hardiness, but must be coupled to the daylength trigger for full resistance to frost (Siminovitch, 1982; Sakai and Larcher, 1987). Initiation of hardening is not a response to physiologically limiting conditions, but occurs in anticipation of these conditions. Photoperiodic control ensures that preparation for dormancy is initiated well in advance of seasonal climatic fluctuation, and is independent of fickle autumn weather. The early/late wood transition, therefore, marks the onset of hardening, and the termination of late wood production corresponds to dormancy and physiologically limiting conditions.

Dormancy in modern temperate plants is most typically a response to seasonal freezing and the accompanying water stress (drought) as water is rendered unavailable, and occurs to prevent tissue damage as temperature drops and ice forms (Li and Sakai, 1978, 1982; Levitt, 1980; Li, 1987; Sakai and Larcher, 1987; Woodward, 1987). Hardening involves energy-demanding biochemical changes that are initiated well in advance of the first frost, typically in mid-summer in temperate regions. Natural selection would eliminate from a region genomes responding much too soon (premature cessation of growth and reduced competitiveness) or too late (damage by early frost) to changing daylength. The initiation of late wood is one of a set of photoperiodically induced changes, so that the early/late wood transition should occur at a similar time of the year within a species in a given region. Variation in the amount of late wood produced from year to year should then correspond to varying climatic conditions of late summer and fall, affecting the length of the growing season (Fritts, 1976; Creber and Chaloner, 1984). Filion et al. (1986), in their studies of black spruce near treeline in Quebec, found that unusually cool temperatures at the end of the growing season correlated with the failure of late wood to develop. These rare "light rings", useful in tree-ring dating at the treeline, represent growth at the extreme limits of tolerance; they are aberrant and are not comparable to the arctic fossil rings.

In the Axel Heiberg fossil forest, extensive development of early wood, and the presence of very little late wood, can be interpreted as indicating near optimal growing conditions throughout the growing season and the abrupt onset of dormancy, as proposed for other high-latitude fossil woods (Parrish and Spicer, 1988). However, the absence of late wood might not be simply an indication of uninhibited growth, but might indicate the lack of an early onset of hardening, or perhaps a lack of hardening at all. This can be interpreted as the lack of a need for hardening and, therefore, an absence of the kinds of winter conditions that would require frost tolerance in plants.

This interpretation puts limits on the minimum winter temperatures experienced by these trees. If the trees were not frost tolerant, minimum winter temperatures could not have remained much below freezing for more than a few hours at a time (Woodward, 1987). It can be argued that resistance to several degrees or more of frost, where supercooling and solute concentration are no longer adequate in themselves, would be so demanding metabolically (Woodward, 1987) that prolonged hardening should be apparent in a substantial production of late wood. The absence of prominent late wood, conversely, indicates the absence of such periods of low temperature.

The discovery of diverse Eocene vertebrate remains on Ellesmere Island (Dawson et al., 1976; West et al., 1977; Estes and Hutchison, 1980; McKenna, 1980) provides solid evidence of mild winters in the Paleogene high latitudes. Modern alligators, for example, cannot withstand prolonged exposure to sub-freezing temperatures (Colbert et al., 1946; Brisbin et al., 1982). However, the concept of a mild but dark winter has been difficult to accept.

Similar interpretations of a mild climate without severe frost and with high equability (equability is defined as the absence of extremes of heat and cold, not lack of temperature variation), have been based on vegetational and paleofloristic analyses of floras comparable to the Buchanan Lake material (Schweitzer, 1974; Wolfe, 1978, 1985; Axelrod, 1984). Axelrod's review of high-latitude floral and faunal evidence, and his discussion of the application of physiognomic and floristic methods of paleoclimatic determination have summarized the arguments for a mild polar paleoclimate. Flora, vegetation, and tree-ring data from the Buchanan Lake Formation are comparable to data for other high-paleolatitude fossils, as discussed by Axelrod, and corroborate evidence of a mild winter in which temperatures remained above freezing or dropped at most to a few degrees below freezing, and then only infrequently. The absence and near-absence of solar radiation during the winter would have resulted in little or no daily fluctuation in temperature, and long-term stability of temperature could be expected.

Regions above the Arctic Circle receive an abundance of light during the growing season and, on a yearly basis, receive an amount that in absolute terms is little different to that received elsewhere on Earth (Creber and Chaloner, 1985), so that the light regime is not a limiting factor for growth. In fact, a proportionately greater amount of light may have been available during the growing season because of the poleward shift in distribution of light from the dormant season to the growing season. The low angle of incidence of light in the high latitudes may have resulted in deep shade on the forest floor (Jefferson, 1982), but ample energy would have been available for canopy plants. The summer temperature regime should be examined in view of this distribution of light. Throughout the period of continuous light, daily fluctuations of temperature in the modern Arctic are low. Similarly, in the past, the summer temperatures would not have been strongly cyclic, and high temperatures would have been maintained over long periods of time. Temperatures on Axel Heiberg Island, recorded during recent field operations, have occasionally gone above 20°C, but typically remain well below this because of the chilling effect of ice caps and a permanently frozen Arctic Ocean. Modern midsummer temperatures, particularly record high temperatures, show remarkably little difference from the tropics to the Arctic Circle, with temperatures commonly reaching the high 20's(°C) or above (Wolfe, 1979; Ruffner and Bair, 1987). With the low latitudinal thermal gradient of the early Tertiary, small diurnal temperature fluctuations, and abundance of light, summer temperatures in the polar regions would be expected to have reached and continued at these levels for extended periods of time.

The interpretation that in the early Tertiary rainfall was abundant throughout the high latitudes, and distributed throughout the year, is supported not only by studies of vegetation but also by the widespread occurrence of thick coal deposits (Meyerhoff, 1983; Orheim, 1983; Ricketts and Embry, 1984). If much of this moisture originated in lower latitudes, this could have represented a significant transfer of latent heat poleward. More importantly, if polar ocean temperatures and oceanic deep water temperatures were as high as 15°C (Shackleton and Kennett, 1975; Kennett, 1977, 1982; Savin, 1977; Frakes, 1979; McKenna, 1983; Shackleton, 1986), then heat energy would have been plentiful in the high latitudes. The presence of a large body of warm water, in an area where land and air were chilled during the dark winter by high net radiative heat loss, may have created dense fog and cloud cover; this may have contributed to temperature moderation through the redistribution of energy as latent heat and a reduction in the rate of atmospheric heat loss.

Evolution of high-latitude vegetation

Although minimum temperatures in the Eocene polar regions may have been above the cardinal minimum temperature of -15°C that limits the development of broad-leaved evergreen vegetation (Woodward, 1987), the long period of winter darkness would have made the survival of such vegetation unlikely. Deciduous broad-leaved vegetation, however, with leaf fall an adaptation to winter darkness, would be expected to have done well under these conditions (Wolfe, 1978, 1980, 1985; Axelrod, 1984). The result would have been the perpetuation of a deciduous, broad-leaved type of vegetation (the Arcto-Tertiary forest) above the Arctic Circle, regardless of thermal seasonality (or lack of it) or mean annual temperature, so long as minimum temperatures were not limiting. Although Wolfe and Poore (1982) documented apparently major Paleocene and Eocene climatic fluctuation, it has not been shown that the vegetation (not necessarily flora) of the high latitudes varied during that time. Vegetation and floristic analyses for the determination of paleoclimatic conditions would be unreliable within such a region. Wolfe and Upchurch (1986), in their study of Cretaceous-Tertiary boundary changes, noted that temperature estimates from this type of deciduous vegetation may be suspect.

The high-latitude floristic changes that occurred between the Eocene and Miocene are best illustrated by the importance of the Pinaceae in Neogene vegetation, as noted above. Because evergreen Pinaceae would be metabolically taxed during a long, dark, mild winter, in much the same way as broad-leaved evergreens, they may have been physiologically excluded from early Tertiary polar lowland floras. However, where winters were for the most part below freezing and metabolic activity could cease, such as at high altitude, evergreen conifers would not be disadvantaged. This may explain the generally poor macrofossil record of Pinaceae in the Paleocene and Eocene, at a time when the pollen record shows abundant and diverse Pinaceae. Enrichment in evergreen Pinaceae is evident in Eocene floras of the northern Rocky Mountains, for which high elevation has been argued (Axelrod, 1966; Wolfe, 1987; Wolfe and Wehr, 1987). Inclusion of Pinaceae within the Buchanan Lake Formation deposits is unlikely to have been a result of a high elevation setting, in view of the depositional model proposed by Ricketts (*this volume*), but rather may indicate that the climate had cooled to a point where evergreen coniferous vegetation was more widespread. It is unlikely, however, given the composition of the Buchanan Lake flora as a whole, that the climate had deteriorated to the point where the lowland Arcto-Tertiary vegetation was threatened.

The onset of Tertiary cooling in Late Eocene time, especially the rapid decline in temperatures associated with the Eocene-Oligocene boundary (Frakes, 1979), would have had a profound effect upon the northern forests (Wolfe, 1978). Firstly, temperature-sensitive taxa would have been eliminated. Secondly, evergreen coniferous vegetation would have become competitive in the lowland habitats (near the sites of deposition). The climatic conditions favouring these changes would have included temperatures that ranged below freezing for much of the dark winter season, possibly comparable to cardinal temperatures. The vegetation changes in the high latitudes may have been abrupt, based on the palynological changes documented in studies by Norris (1982, 1986). The replacement of deciduous forest by evergreen coniferous forest in the high latitudes, as a result of declining winter temperatures, does not conform to Wolfe's (1987) proposal that this change was related to declining summer temperatures, although Wolfe (1978, 1980) did recognize a post-Eocene decline in equability.

High-latitude Tertiary climatic change, whether a product of changing "greenhouse" conditions or decreasing efficiency of global heat transport mechanisms (Barron, 1985), would have been most strongly felt during the winter, when net radiative loss was the greatest. In the absence of the cooling effect of permanent ice, solar radiation is more than adequate to warm the Arctic, but, during winter darkness, temperatures would be determined more directly by the rate of heat transfer to the poles. Declining polar transport of heat during the Tertiary would have caused declining winter temperatures and progressive shortening of the growing season, but may have had less influence on warm month mean temperature.

Initiation of northern hemispheric glaciation in the Late Pliocene (Frakes, 1979) would have brought an end to high polar summer temperatures and led to the development of a treeline.

CONCLUSIONS

The Buchanan Lake Formation of northeastern Axel Heiberg Island preserves a predominantly deciduous Arcto-Tertiary vegetation, including many undescribed taxa. Floristic composition, vegetation characteristics, and wood anatomy support the interpretation that the Eocene polar climate was mild and equable, and without appreciable frost in the winters. However, the occurrence in the fossil flora of diverse, although typically rare Pinaceae, with early occurrences of a number of genera, indicates expansion of coniferous evergreen vegetation during Buchanan Lake time; this may indicate the onset of climatic deterioration.

Declining winter temperatures and decreased equability probably were the most significant aspects of Tertiary climatic deterioration in terms of high-latitude vegetation. In the past, sufficient heat was transported poleward to prevent deep chilling during the dark winter period. In the modern polar regime, heat transfer to the very high latitudes is minimal and the grip of winter is never lost.

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PLATES 1 TO 4

PLATE 1

Figure 1. View of the Axel Heiberg Island site showing an exposure of one of the many fossil forests. The tree stumps shown in this photograph are over one metre in diameter, and are among the largest preserved at the site. All materials figured are derived from this and nearby sites, unless otherwise noted.

Figure 2. Cross-section of fossil wood, showing the wide annual increment of growth and little indication of late wood development. Growth was from left to right. (A) = beginning of the increment, (B) = end of the increment. US400-4569, x25.

Figure 3. Cross-section of the wood of the Douglas fir, *Pseudotsuga menziesii*, a western Canadian conifer, showing well developed late wood of the annual increments, x50.


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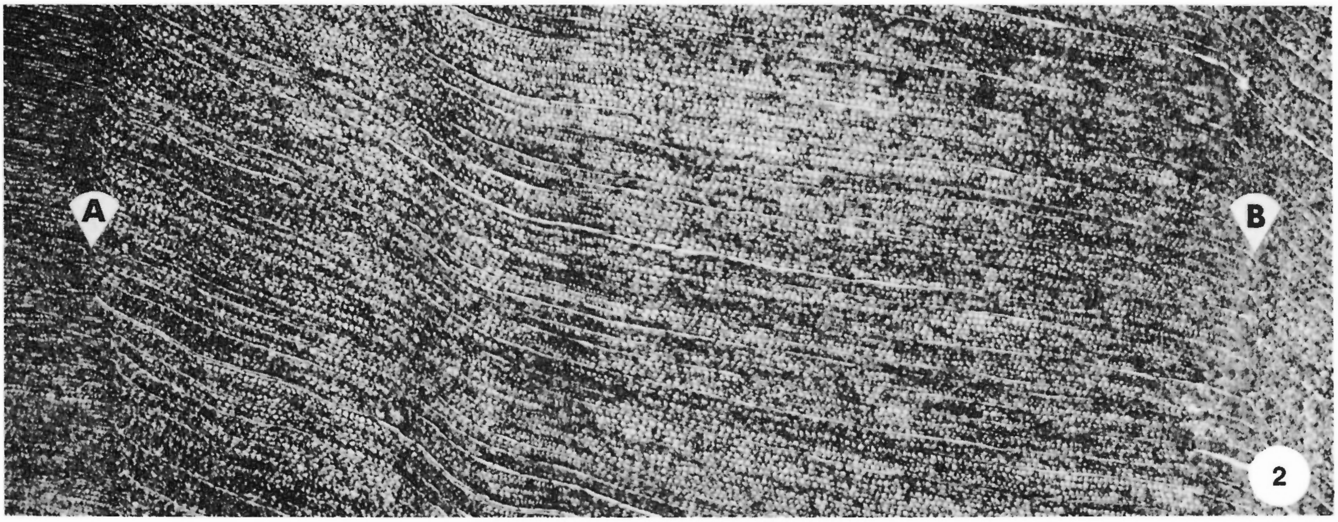
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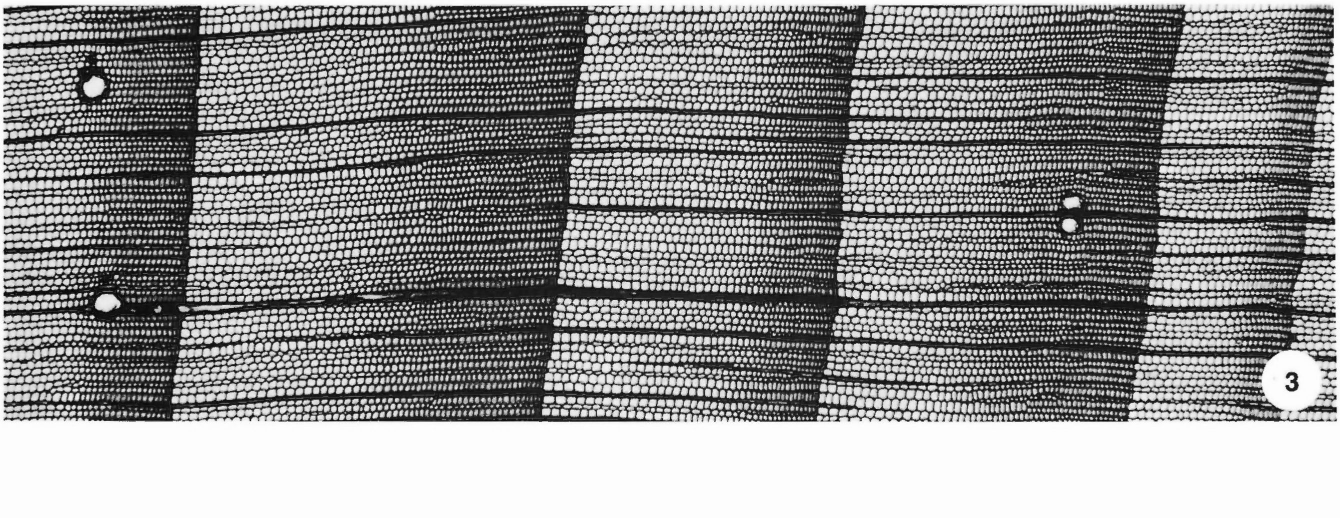
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PLATE 2

Figure 1. One of the many layers of leaf litter preserved at the Axel Heiberg Island site. The compact remains of the forest floor litter can be removed in bulk.

Figure 2. Leafy twigs of the dawn redwood, *Metasequoia occidentalis* (Newberry) Chaney, recovered from leaf litter, x1.5.

Figure 3. Leafy twigs of the swamp cypress, *Glyptostrobus nordenskioldii* (Heer) Brown, from leaf litter, x1.5.

Figure 4. Cones of *Glyptostrobus nordenskioldii* from leaf litter. US185-4570 (upper), US185-4576 (lower), x1.5.

Figure 5. Cones of *Metasequoia occidentalis* from leaf litter. US220-4571, x1.5.

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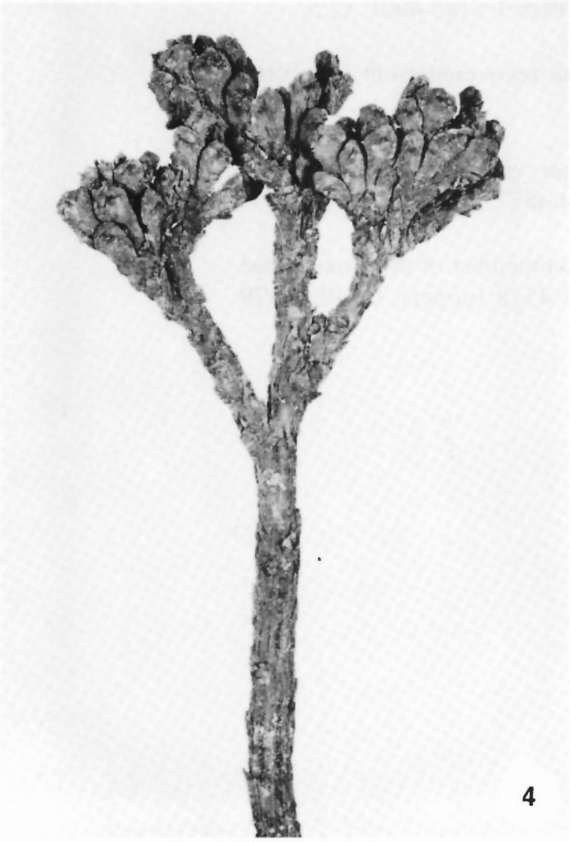

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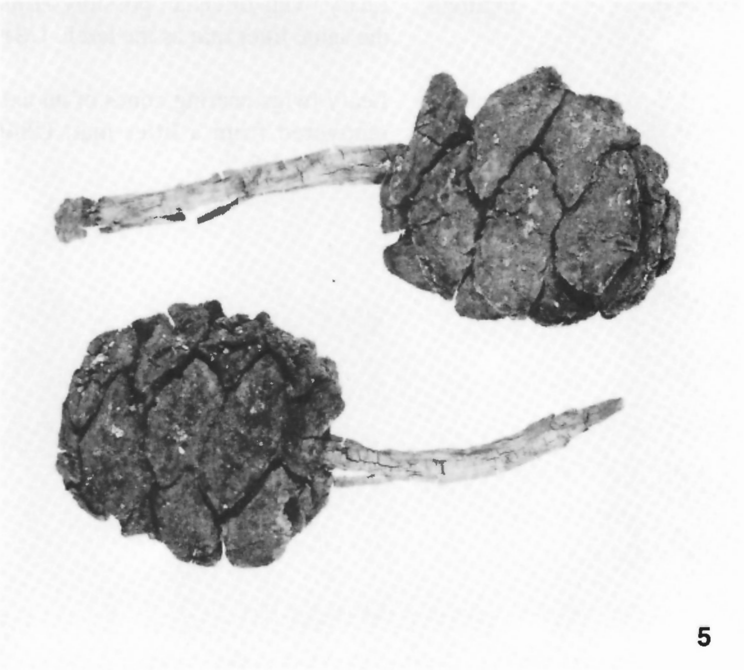

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PLATE 3

Figure 1. Large-coned spruce, *Picea* sp., cone recovered from organic debris preserved within fluvial sands. US120-4572, x1.5.

Figure 2. Small-coned spruce, *Picea* sp., cone recovered from fluvial sands. US120-4573, x1.5.

Figure 3. Cone of pine, *Pinus* sp., recovered from fluvial sands. US115-4574, x1.5.

Figure 4. Cone of larch, *Larix* sp., attached to brachioblast of twig. Remains of larch are a rare occurrence in leaf litters. US185-4001, x2.5.

Figure 5. Leaf fascicles of a three-needle pine recovered from leaf litter. US401-4575 (all), x1.

Figure 6. Leafy twigs of cedar, possibly *?Chamaecyparis* sp., recovered from the same litter mat as the larch. US184-4577, x2.

Figure 7. Leafy twigs bearing cones of an extinct member of the Taxodiaceae recovered from a litter mat. US402-4578 (upper), US402-4579 (lower), x1.5.


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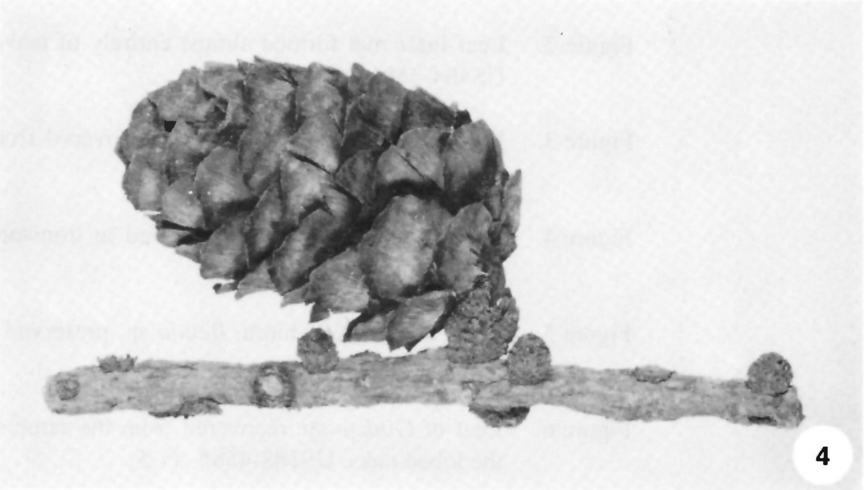
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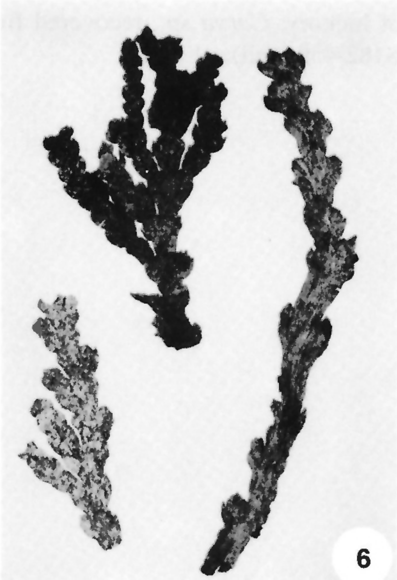
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PLATE 4

Figure 1. Leaf of a lobed oak, *Quercus* sp., recovered from siltstones exposed near the fossil forest site. US188-4580, x1.

Figure 2. Leaf-litter mat formed almost entirely of leaves of alder, *Alnus* sp. US404-4581 (all), x1.

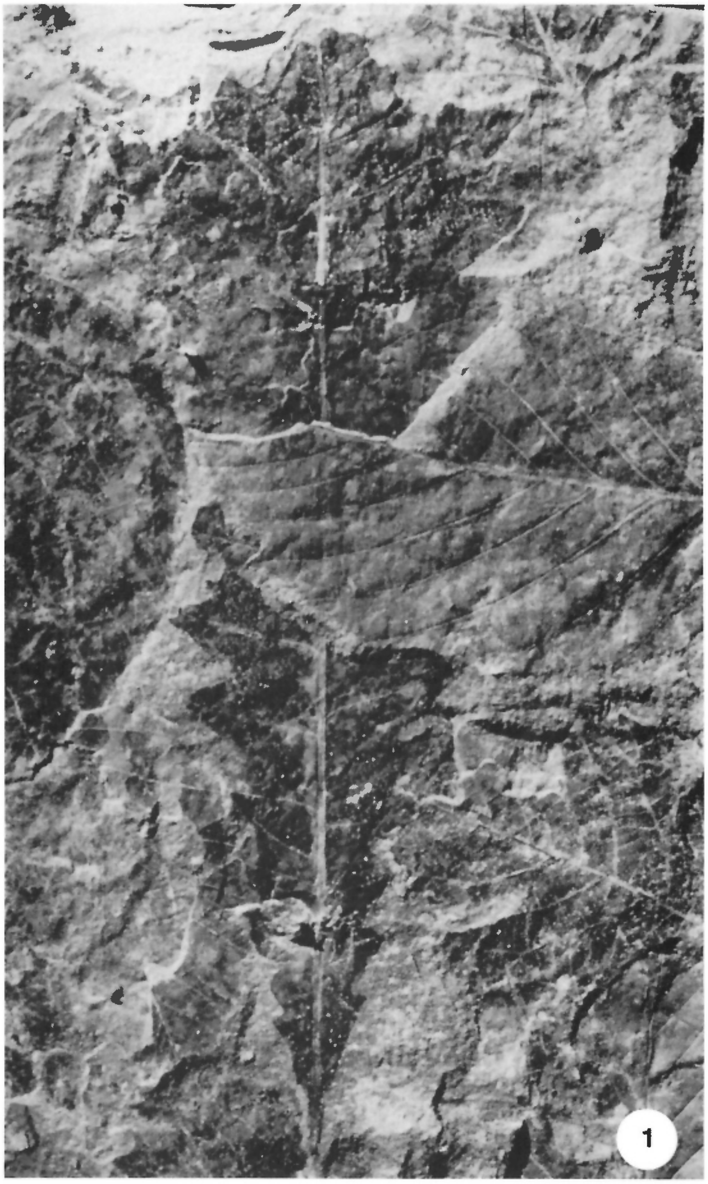
Figure 3. Fruits of "*Cercidiphyllum*" sp. recovered from leaf litter. US405-4582 (all), x1.5.

Figure 4. Infructescences of alder preserved in ironstone at the fossil forest site. US187-4583, x2.

Figure 5. Infructescences of birch, *Betula* sp. preserved in leaf litter. US220-4584, x2.5.

Figure 6. Leaf of *Ginkgo* sp. recovered from the same siltstones that contain the lobed oaks. US188-4585, x1.5.

Figure 7. Fruits of hickory, *Carya* sp., recovered from fine grained fluvial silts. US182-4586 (all), x1.5.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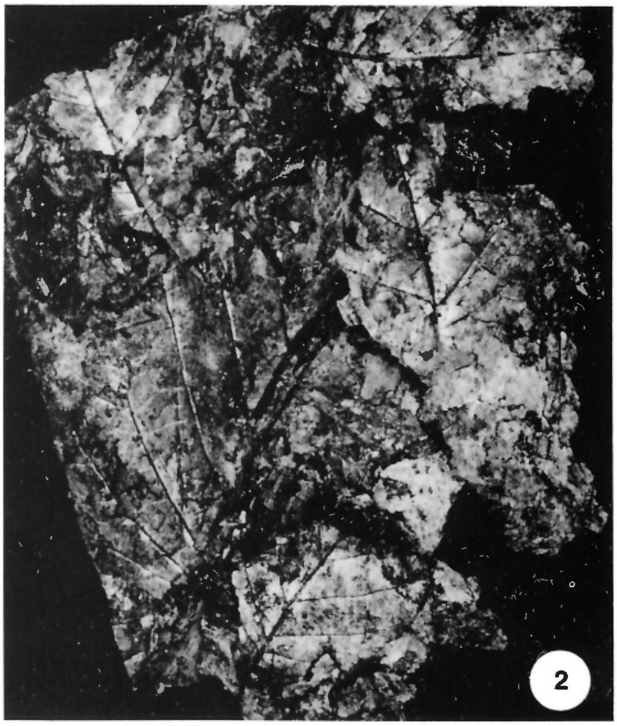
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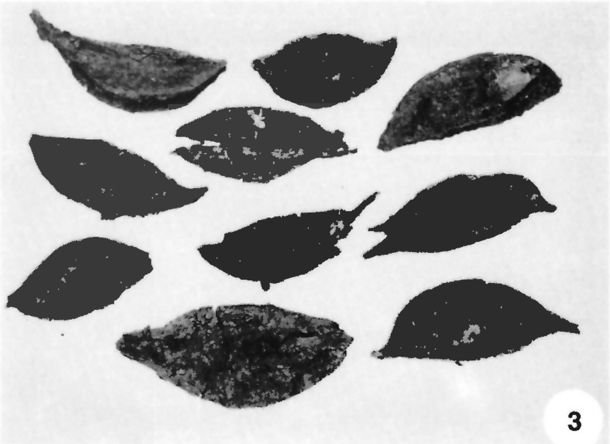
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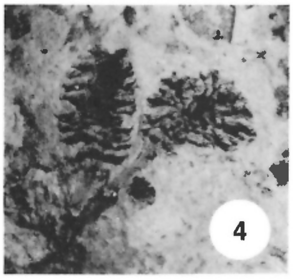
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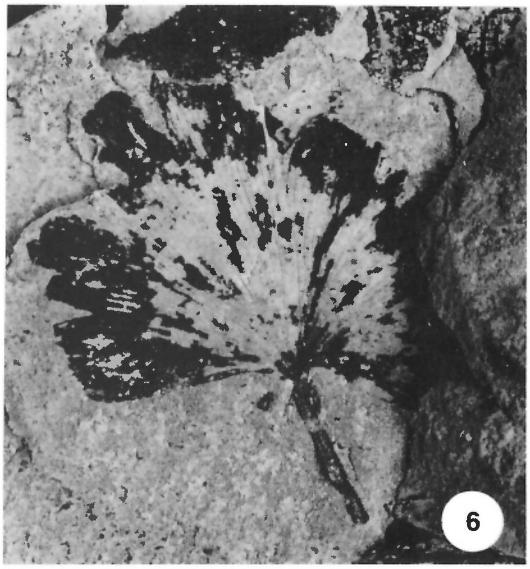
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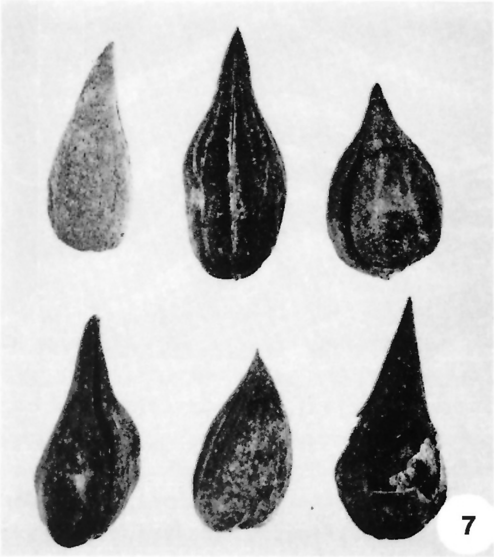
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EARLY TERTIARY *LARIX* FROM THE BUCHANAN LAKE FORMATION, CANADIAN ARCTIC ARCHIPELAGO, AND A CONSIDERATION OF THE PHYTOGEOGRAPHY OF THE GENUS

Ben A. LePage[1] and James F. Basinger[1]

LePage, B.A. and Basinger, J.F., 1991. Early Tertiary ***Larix*** *from the Buchanan Lake Formation, Canadian Arctic Archipelago, and a consideration of the phytogeography of the genus. In Tertiary Fossil Forests of the Geodetic Hills, Axel Heiberg Island, Arctic Archipelago, R.L. Christie and N.J. McMillan (eds.); Geological Survey of Canada, Bulletin 403, p. 67-82.*

Abstract

Exquisitely preserved fossil forests of Eocene age have been recently discovered on Axel Heiberg Island at 80°N in the Canadian Arctic Archipelago. In situ stumps and leaf litters of the forest floors, with allochthonous accumulations of plant debris, occur within fine grained floodplain deposits of the Buchanan Lake Formation, Eureka Sound Group. Within some of the autochthonous litter mats are the earliest known fertile and vegetative remains of *Larix*. Extant *Larix* form two natural groups: the widely distributed short-bracted forms, and the more restricted alpine long-bracted forms. The Buchanan Lake *Larix*, together with all other reports of fossil *Larix*, indicate that the short-bracted group has a long history as a component of high-latitude coniferous forests. The morphologically and ecologically distinct long-bracted species as yet lack a fossil record. The spatial and temporal distribution of fossil *Larix* indicates that the genus was dispersed throughout the Canadian Arctic, Beringia, and northeastern Asia in the late Paleogene and early Neogene, but reached Europe only in the Pliocene. A review of paleogeographic evolution of the Northern Hemisphere reveals that the late Paleogene destruction of the North Atlantic dispersal routes and the relatively low paleolatitude of Europe may have prevented migration of *Larix* eastward from North America to Europe. Physical continuity between Europe and Asia was established in the late Paleogene following regression of the Turgai Strait, but dispersal of *Larix* into Europe was apparently delayed until the Pliocene due to the climate.

Résumé

Des forêts fossiles d'âge éocène, conservées de façon remarquable, ont récemment été découvertes dans l'île Axel Heiberg à 80°N dans le Haut-Arctique canadien. Des souches et litières trouvées in situ, qui faisaient partie du sol forestier, accompagnées d'accumulations allochtones de débris végétaux, existent dans des sédiments de granulométrie fine déposés sur une plaine inondable, dépôts appartenant à la formation de Buchanan Lake dans le groupe d'Eureka Sound. À l'intérieur de quelques-unes des accumulations autochtones de litière, ont été trouvés les plus anciens restes connus d'organes reproductifs et végétatifs de *Larix*. Les formes actuelles de *Larix* constituent deux groupes naturels: les formes à bractées courtes, de large distribution, et les formes alpines à longues bractées, de distribution plus restreinte. Le *Larix* de la formation de Buchanan Lake, en même temps que tous les autres *Larix* fossiles signalés, indique que le groupe à bractées courtes forme depuis longtemps une composante des forêts de conifères des latitudes élevées. L'espèce à bractées longues, morphologiquement et écologiquement distincte, est pour l'instant absente de la colonne stratigraphique. La distribution spatiale et temporelle de la forme *Larix* fossile indique que ce genre a été dispersé dans tout l'Arctique canadien, dans Beringia, et l'Asie du Nord-Est au Paléogène supérieur et au Néogène inférieur, mais n'a atteint l'Europe qu'au Pliocène. Un examen de l'évolution paléogéographique de l'hémisphère nord révèle que la destruction au Paléogène supérieur des voies de dispersion nord-atlantiques et la paléolatitude relativement basse de l'Europe ont peut-être empêché la migration de *Larix* vers l'est entre l'Amérique du Nord et l'Europe. La continuité physique de l'Europe et de l'Asie s'est établie au Paléogène supérieur, en raison de la régression survenue dans le détroit de Turgai, mais la dispersion de *Larix* en Europe a apparemment été retardée jusqu'au Pliocène pour des raisons climatiques.

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[1] Department of Geological Sciences, University of Saskatchewan, Saskatoon, Saskatchewan S7N 0W0

INTRODUCTION

The genus *Larix* Miller (1754), commonly referred to as larches or tamaracks, is widely distributed across North America, Asia, and Europe and forms a prominent component of northern boreal forests, but little is known about the evolutionary and phytogeographic history of the genus. Prior to the investigation of the Eocene fossil forests of the Buchanan Lake Formation on Axel Heiberg Island, the fossil record for *Larix* could be extended with confidence only into the Oligocene (Miller, 1977, 1988). The presence of fertile (seed cones) and vegetative (twigs and needles) *Larix* remains within the Buchanan Lake leaf-litter mats is significant, for it provides evidence of pre-Oligocene evolution for the genus.

The occurrence of *Larix* in the Canadian High Arctic indicates that the genus existed in the high latitudes during the Eocene and may have dispersed into the mid-latitudes of North America and Eurasia as polar climates deteriorated. The onset of Tertiary climatic deterioration in the Eocene (Wolfe, 1978) correlates with the tectonic evolution of the Norwegian-Greenland Seas and the emergence of extensive boreal forests dominated by evergreen conifers throughout the high paleolatitudes of the Northern Hemisphere.

High-latitude land bridges (the DeGeer and Beringian routes) have existed intermittently between North America and Eurasia during much of the Cenozoic and have facilitated intercontinental dispersal of both flora and fauna (Chaney, 1940, 1947; Hopkins, 1967; McKenna, 1972a,b; Wolfe, 1972, 1975; Tiffney, 1985a,b). The spatial and temporal distributions of fossil *Larix* throughout the Northern Hemisphere indicate that some species may have made free use of the Beringian corridor throughout much of the Cenozoic. Warmer climates associated with the North Atlantic region during the early Cenozoic, and the destruction of the North Atlantic Routes at about the time of the Eocene/Oligocene boundary evidently precluded *Larix* from dispersing eastward into Fennoscandia, Europe, and Asia from the Canadian Arctic.

In this paper we examine the distributions of extinct and extant *Larix* and, together with information concerning the availability of land bridges, present an interpretation of the Cenozoic phytogeography of *Larix*.

MATERIALS

Geological setting

The Eureka Sound Group is a regionally distinctive sedimentary sequence occurring within the Sverdrup Basin of the Queen Elizabeth Islands, Arctic Canada. Miall (1986) and Ricketts (1986) independently proposed subdivision of this sedimentary sequence. Ricketts (1986, *this volume*) has examined the deposits containing the fossil forests during the present study, and his nomenclature has been adopted in this paper. Four formations, from the Middle or Late Campanian to the Middle or Late Eocene, are recognized: Expedition Formation; Strand Bay Formation; Iceberg Bay Formation; and the uppermost and youngest of the group, the Buchanan Lake Formation.

Deposition of the Buchanan Lake Formation occurred during the last phase of the Eurekan Orogeny (Tozer, 1960; Ricketts, 1986, 1987, *this volume*; Ricketts and McIntyre, 1986). Initiation of seafloor spreading between Norway and Greenland (Talwani and Eldholm, 1977) and changes in the rate of seafloor spreading in the Labrador Sea (Srivastava, 1978) resulted in regional deformation, including uplift of the Princess Margaret Arch, and crustal shortening throughout this region (Kristoffersen and Talwani, 1977). Thrusting of older Mesozoic and Paleozoic bedrock along the Stolz Thrust Zone on eastern Axel Heiberg Island resulted in the synorogenic deposition of gravels, sands, and silts by braided-river systems and debris flows in the basin forming at the foot of the rising Princess Margaret Mountains (Ricketts, 1986; Ricketts and McIntyre, 1986).

Within the Buchanan Lake Formation, Ricketts (1986, *this volume*) recognized four lithologically distinctive units: 1) a lower coal-bearing member consisting of interbedded lithic sandstone, mudstone, and lignite; 2) a massive conglomeratic alluvial deposit, which may be the upslope equivalent of the alluvial plain, lower coal-bearing member; 3) an interbedded conglomerate, sandstone, and grey mudstone/siltstone member; and 4) an upper lignitic coal-bearing member consisting of interbedded lithic sandstone, mudstone/siltstone, and lignite arranged in fining-upward sequences. The autochthonous leaf-litter mats and in situ woody stumps, as well as allochthonous deposits yielding cones, seeds, and leaf compressions, found at the fossil forest site occur in the upper lignitic coal-bearing member (Ricketts, 1986, *this volume*; Ricketts and McIntyre, 1986; Basinger, *this volume;* Francis, *this volume*).

The age of the sediments bearing the fossil forests and the remains of *Larix* has been determined as Middle or possibly Late Eocene (Ricketts and McIntyre, 1986; McIntyre, *this volume*).

Floral setting

The fossil forest site is located east of the Geodetic Hills at 79°55'N, 88°58'W on eastern Axel Heiberg Island (Fig. 1). Irving and Wynne (*this volume*) have determined that the paleolatitude of the site was at most a few degrees south of its present position, and that Axel Heiberg Island was therefore well above the Arctic Circle when the forests were alive. Extensive late Cenozoic or Quaternary erosion has resulted in widespread exposure of the forest-bearing beds. Exquisitely preserved remains of *Larix* (Fig. 2) were obtained from autochthonous leaf-litter mats in the middle portion of the upper lignitic coal-bearing member at the fossil forest site. The *Larix* remains occur in the upper part of the section containing the greatest concentration of litter mats (as discussed by Basinger, *this volume*). The leaf-litter mats represent the ancient forest floors of poorly drained, low energy, meander-plain swamps (Ricketts, 1986, *this volume*). The constantly wet, anaerobic environment allowed mummification of the detrital material. The collection of *Larix* fossils currently consists of approximately 200 seed cones, 15 intact needle fascicles, 5 leading shoots, and numerous long shoots with brachioblasts (short shoots).

Megafloral remains recovered from the same leaf-litter mats that contain *Larix* include: the dominant conifers *Metasequoia* (dawn redwood), and *Glyptostrobus* (swamp cypress); and uncommon *Picea* (spruce), ?*Pseudolarix* (golden larch), ?*Abies* (fir), *Pinus* (pine), *Betula* (birch), *Carya* (hickory), ?*Chamaecyparis* (false cedar), ?*Tsuga* (hemlock), and *Osmunda* (royal fern) (see also Basinger, 1986, *this volume*; Basinger et al., 1988). The scarcity of *Larix* relative to other taxa suggests that the larches were uncommon constituents within a Taxodiaceae dominated swamp-forest community.

DISCUSSION

The genus *Larix*

Species of *Larix* (extant and extinct) are some of the few conifers that are both deciduous and possess dimorphic branching systems and foliar arrangements. The main branches, or long shoots (Fig. 2a), bear smaller short shoots or brachioblasts that, in turn, bear either erect seed cones (Fig. 2a)

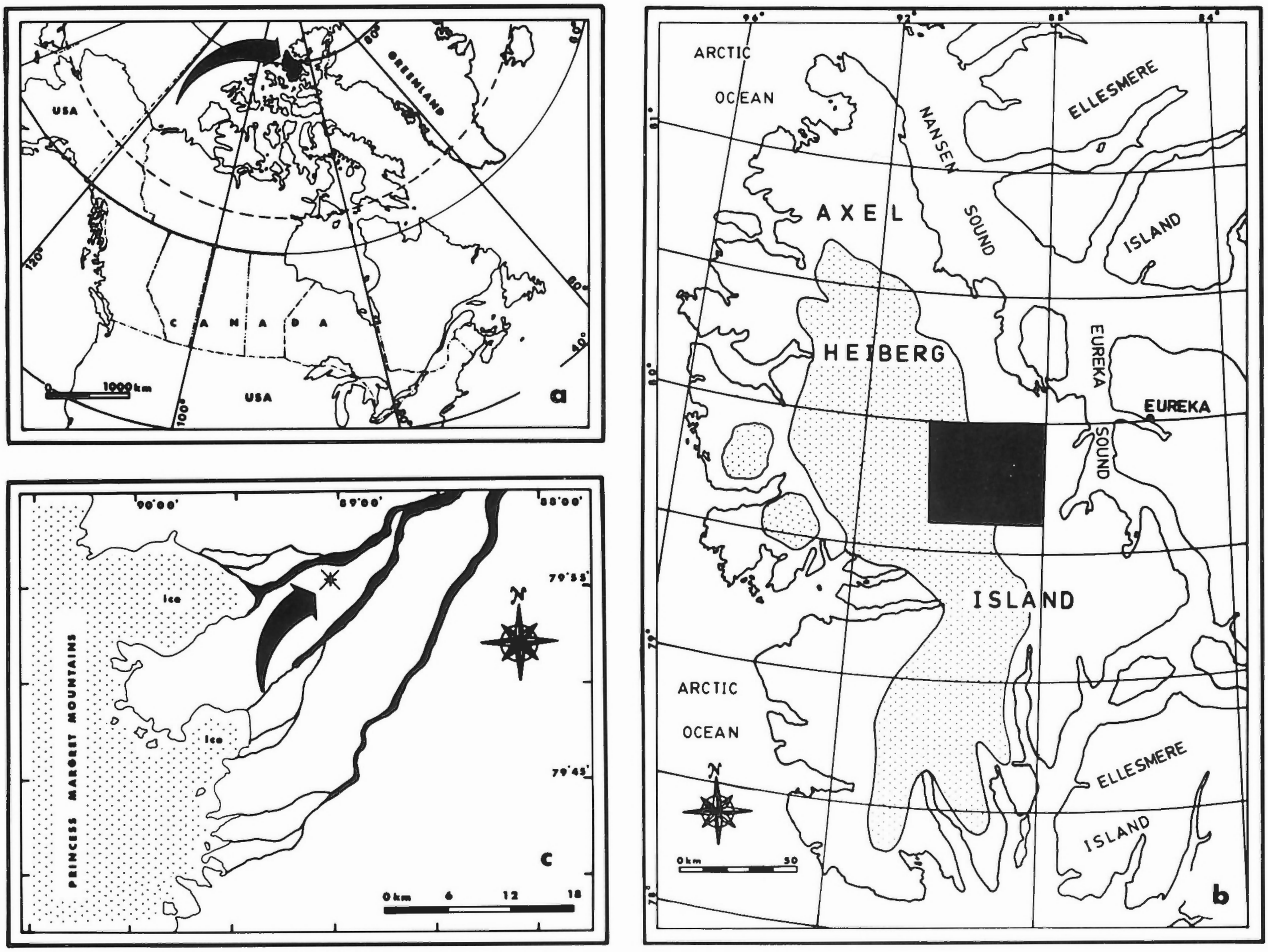

Figure 1. *Map of study area: a, map of Canada showing the location of Axel Heiberg Island; b, detailed map of Axel Heiberg Island showing the location of the fossil forest area; and c, detailed area map showing the fossil site. Stippled area = ice.*

or small tufts of needles called fascicles (Fig. 2b). The second type of foliar arrangement occurs on the new growth shoots, or leading shoots, of the tree (Fig. 2c), where needles are arranged helically along the length of the leading shoot, rather than in fascicles.

Extant distribution

The genus *Larix* currently consists of ten species and three varieties (Ostenfeld and Larsen, 1930a,b); three species are endemic to North America, with the remainder occurring

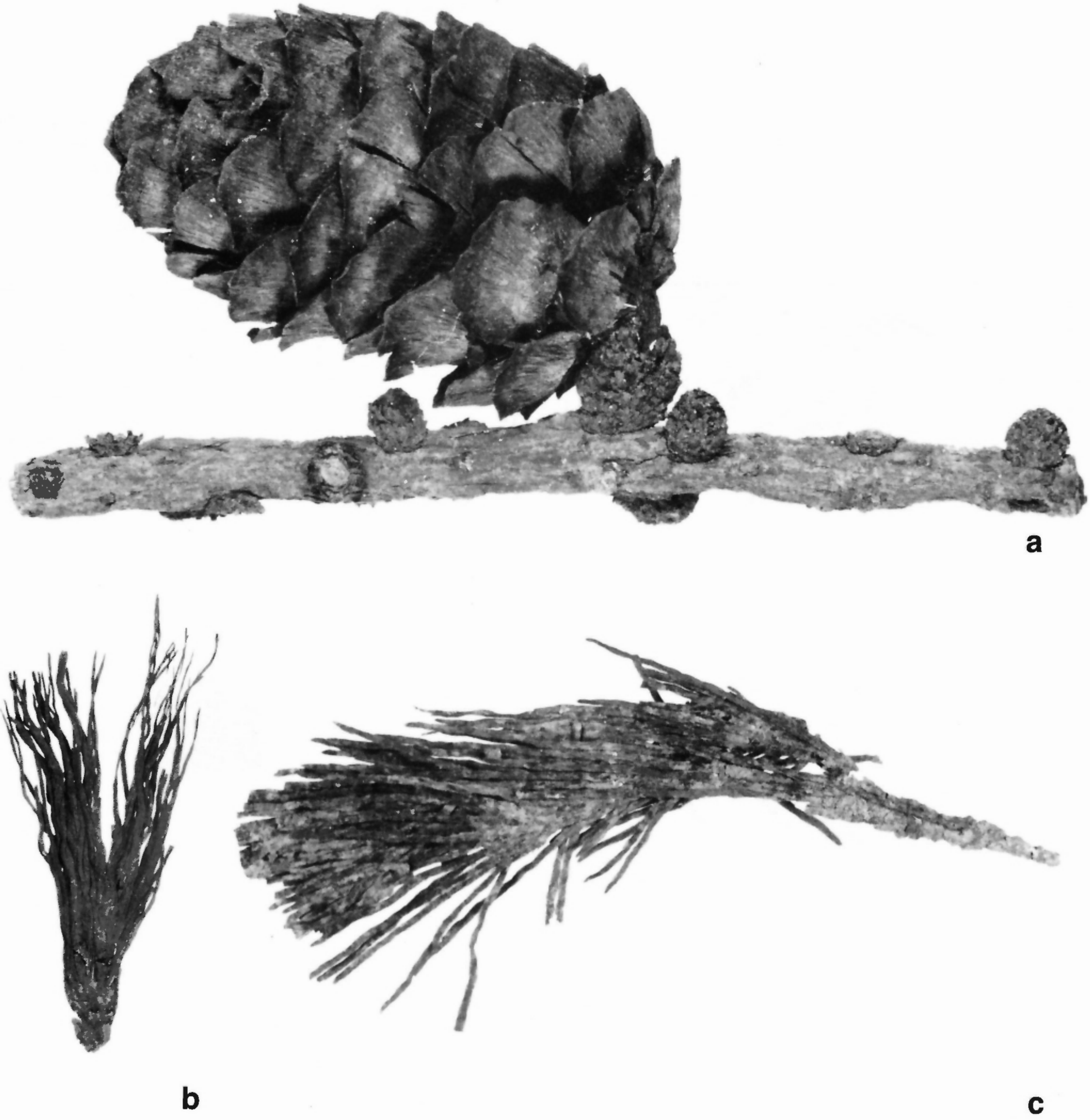

Figure 2. *a, cone of **Larix** attached to a brachioblast (short shoot) and borne on a long shoot. US185-4003, x4. b, fascicular needles of fossil **Larix**. US185-4001, x1.5. c, leading shoot of fossil **Larix** showing the arrangement of the needles. US220-4000, x1.6.*

throughout Asia and Europe (Fig. 3). *Larix laricina* (Du Roi) K. Koch of North America and *L. gmelini* Turcz. and *L. sibirica* Ldb. of Asia have extensive, continental distributions, while the remaining species are geographically and possibly environmentally restricted to mountainous regions (Ostenfeld and Larsen, 1930a,b; Little, 1971; Takhtajan 1986). The species of *Larix* occupy a variety of geographic provinces from alpine to temperate lowland.

Distribution of extant species of *Larix* provides evidence of three main patterns of dispersal. In North America, in addition to the widespread distribution of *L. laricina*, there apparently was a southwestward dispersal along the Cordillera, so that today two of the three living species are found with limited distributions along the eastern Rocky Mountains of Canada and the United States. Within Asia, two distinct patterns of dispersal emerge: the first along the eastern margin of Asia and into central China; the second, in northern Asia and eastern and central Europe. These dispersal patterns will be referred to as: 1) the North American, 2) the Southern Asian, and 3) the Trans-Eurasian.

In North America, *L. occidentalis* Nutt. and *L. lyallii* Parl. are found growing in the Rocky Mountains of southern Alberta and British Columbia, northern Washington, western Montana, and Idaho. The widespread *L. laricina* occurs from the eastern margin of the Rocky Mountains as far north as Alaska, across most of Canada, and south into the Great Lakes region of the United States.

The Southern Asian pattern includes several members of the genus. *Larix gmelini* ranges from northeastern Siberia

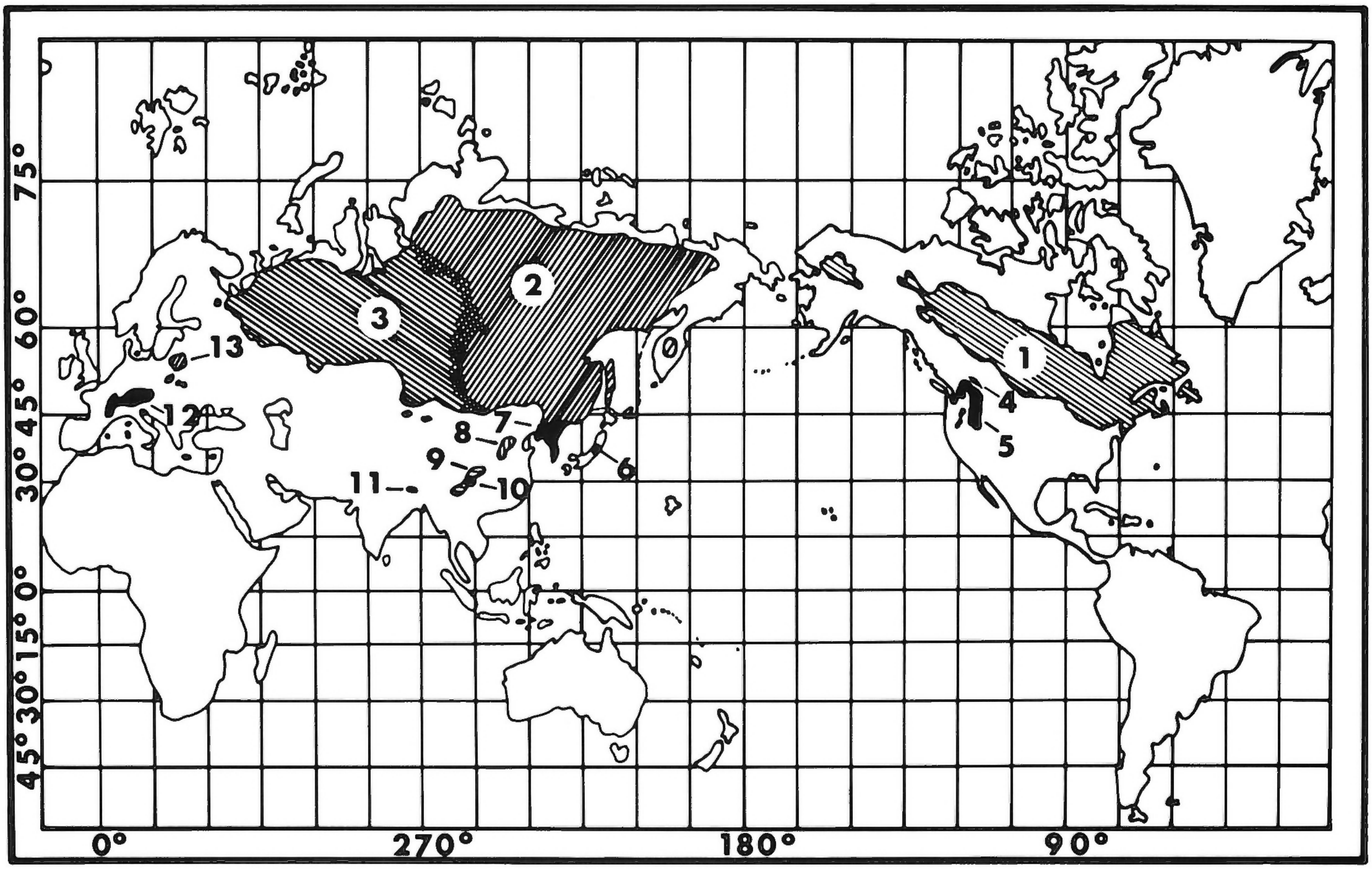

Figure 3. Generalized map of the world showing the global distribution of extant species of ***Larix****. Legend of scientific and common names: 1,* ***Larix laricina*** *(Du Roi) K. Koch (Tamarack); 2,* ***L. gmelini*** *Turcz. (Dahurian Larch); 3,* ***L. sibirica*** *Ldb. (Siberian Larch); 4,* ***L. occidentalis*** *Nuttall (Western Larch); 5,* ***L. lyallii*** *Parlatore (Subalpine Larch); 6,* ***L. kaempferi*** *(Lambert) Sargent (Japanese Larch); 7,* ***L. gmelini*** *var.* ***olgensis*** *(Mayr) Ostenfeld & Syrach Larsen (Kurile Larch); 8,* ***L. gmelini*** *var.* ***principis-rupprechtii*** *(Mayr) Ostf. & Syrach L. (Prince Rupprech Larch); 9,* ***L. potanini*** *Batalin (Chinese Larch); 10,* ***L. griffithiana*** *(L. & G.) Carriere (Himalayan Larch); 11,* ***L. mastersiana*** *Rehder and Wilson (Masters Larch); 12,* ***L. decidua*** *Miller (European Larch); 13,* ***L. decidua*** *var.* ***polonica*** *(Raciborski) Ostf. & Syrach L. (Polish Larch). (Map is adapted from Ostenfeld and Larsen, 1930a,b; Komarov, 1934; Szafer, 1946, 1966; Dallimore and Jackson, 1948; Li, 1952; Little, 1971; Zoltai, 1973; Scoggan, 1978; Silba, 1986; and Takhtajan, 1986.)*

southward along the southeastern coast of Asia, where small regional populations of *L. gmelini* var. *olgensis* (Mayr) Ostf. & Syrach L., *L. gmelini* var. *principis-rupprechtii* (Mayr) Ostf. & Syrach L., and *L. kaempferi* (Lambert) Sargent are found. From Korea this pattern extends southwest into the mountainous regions of southern China, where *L. mastersiana* Rehder & Wilson and *L. potanini* Batalin occur in the same region, although at different altitudes. Farther west, into the Himalayas, *L. griffithiana* (L. & G.) Carr. occurs as a treeline species.

Within the Trans-Eurasian pattern, *L. gmelini* (eastern) and *L. sibirica* (western) occur as prominent components of the boreal forest. Where the two distributions overlap in the central Soviet Union, hybridization occurs (Ostenfeld and Larsen, 1930a). *Larix sibirica* occurs as far west as Moscow, while *L. decidua* Miller is found in the mountainous regions of central Europe. An apparent hybrid between *L. sibirica* and *L. decidua*, *L. decidua* var. *polonica* (Raciborski) Ostf. & Syrach L., indicates that these two species once were more extensively distributed.

The extant species of *Larix* appear to fall into two major groups. The widespread boreal species, *L. laricina*, *L. sibirica*, *L. gmelini* (including the more restricted varieties of this species), and the more restricted European species, *L. decidua* and *L. decidua* var. *polonica*, possess short, nonemergent bracts and form a natural group. The remaining species, all restricted to montane regions of southeastern and eastern Asia and western North America, possess long, exserted bracts and form a second natural group. In order to consider questions regarding phylogenetic relationships and the phytogeographic history of *Larix*, it is necessary to examine the fossil record.

Fossil distribution

The discovery of short-bracted *Larix* on Axel Heiberg Island extends the fossil record of *Larix* to the Eocene. *Larix* is also recognized with confidence in Tertiary beds in Alaska, Northern Canada, Japan, Poland, and Russia (see Table 1 and Fig. 4). Reports of *Larix* of an Upper Oligocene age (Blokhina, 1984, 1985) and accounts of mass burial of *Larix* in the Miocene of Russia (Piemenov, 1986) demonstrate that *Larix* was well established in northeastern Asia by the end of the Oligocene. Tanai and Onoe (1961) also reported that *Larix* was present in Japan by the Late Miocene. Together with the reports of *Larix* fossils from northern North America (see Table 1), it appears that the distribution of *Larix* may have been high-latitude circumboreal throughout middle to late Cenozoic time. But further, the fossil distribution shown in Figure 4 (for the ages of these fossil localities see Table 1) indicates an apparent absence of *Larix* from the Tertiary deposits of Greenland, Svalbard, and Iceland, and a relatively late (Pliocene) appearance of *Larix* in western Europe (Szafer, 1946, 1966). The Axel Heiberg fossil forests have yielded the oldest and most morphologically complete fossil remains of *Larix*, indicating that the genus was present at least in the Canadian High Arctic in pre-Oligocene time.

To date, remains of long-bracted *Larix* species have yet to be found. Extant long-bracted forms occur only in montane regions, and many are highly restricted in distribution and appear relictual. The modern pattern of distribution and the occurrence of species on both sides of the Pacific indicate that the long-bracted species are an old group. Past distribution of the long-bracted taxa is difficult to ascertain, except that amphi-Pacific distribution indicates Tertiary dispersal across Beringia. Their absence in the fossil record may be related to their adaptation to alpine habitats, where chance of entry into the fossil record is remote. Consequently, discussion of *Larix* phytogeography will be limited to the short-bracted species.

It is interesting that all of the known fossil remains of *Larix*, including those from Axel Heiberg Island, show affinity with the group of short-bracted species that are now widely distributed in the boreal forests. However, modern distributions of *Larix* show that the short-bracted species, as compared with the long-bracted species, typically occupy habitats at lower altitudes where the chance of preservation is greater. Therefore, the fossil record of *Larix* suggests that a taphonomic bias may exist for species of the lowland and swamp habitats and this would account for the over representation of the short-bracted species in the fossil record.

Extant short-bracted taxa include widely distributed members of the boreal forest, and it appears from the fossil record that they have occupied similar niches since at least the Oligocene. The single locality of Eocene *Larix* on Axel Heiberg Island is inadequate to extend this generalization further. However, the presence of *Larix* in the Canadian Arctic during the Eocene does allow speculation on possible Cenozoic dispersal of short-bracted *Larix* in the high latitudes. The North American and Trans-Eurasian patterns of distribution of short-bracted forms could have resulted from dispersal across either the North Atlantic or Beringian land bridges. It is, therefore, necessary to evaluate the availability of these routes in the phytogeographic interpretation of past and present distribution of these taxa. Three routes may have been available for boreal intercontinental plant dispersal during Paleocene and Eocene time: the DeGeer and the Thulian routes between North America and Fennoscandia; and the Beringian Route between North America and eastern Asia (Fig. 5).

The DeGeer Route

The DeGeer Route (McKenna, 1972a) was an important land bridge that extended between North America and Fennoscandia (the ancient Scandinavian Peninsula) throughout the Paleocene and Eocene. The importance of the DeGeer

TABLE 1

Fossil record of the genus *Larix*

Location/locality	Fossil type[1]	Name	Date	Author
Axel Heiberg Island	C, V	*Larix* sp.	Eocene	*(this paper)*
Geodetic Hills"a"				
Geodetic Hills"b"	C	*Larix* sp.	Eocene	Bustin (1982)
Meighen Island	V	*Larix* sp.	Pliocene	Matthews (1987a,b)
Prince Patrick Island				
Green Bay Section	unreported	*Larix* sp	Miocene	Matthews (1987b)
Duvaney Section	unreported	*Larix* sp.	Miocene	Matthews (1987b)
Banks Island				
Ballast Brook	W	*Laricioxylon occidentaloides* Roy and Hills	Miocene	Roy and Hills (1972); Matthews et al. (1986)
Ballast Brook	V	*Larix* sp.	Miocene	Matthews (1987b)
Duck Hawk Bluffs	C	*Larix* cf. *omoloica* Dorof.	Miocene	Hills (1975); Matthews (1987a)
Duck Hawk Bluffs	C, V, W, P	*Larix* sp.	Pleistocene	Matthews et al. (1986)
Worth Point	W	*Larix laricina*	Pleistocene	Kuc (1974); Matthews et al. (1986)
Devon Island				
Haughton Astrobleme	C, V, P	*Larix* sp.	Miocene	Omar et al. (1987)
Yukon Territory				
Bluefish Exposure	V	*Larix* sp.	Neogene	Matthews (1987b)
Twelvemile Bluff (HH 228)	C, V, W, P	*Larix minuta*	Pliocene	Matthews (1987b); Schweger (1987)
Henderson Bluff	C, P	*Larix minuta*	Pliocene	Schweger (1987)
Bluefish River (HH75-24)	P	*Larix* sp.	Miocene	Schweger (1987)
Alaska				
Lava Camp Mine	C, V	*Larix* sp	Pleistocene	Hopkins et al. (1971); Matthews (1987b)
Cone Bluff	unreported	*Larix* sp.	Pleistocene	Matthews (1987b)
North Slope	V	*Larix* sp.	Pliocene	Matthews (1987b)
Russia				
Siziman Bay	W	*Laricoxylon jarmolenkoi* Blokhina	Oligocene	Blokhina (1985)
Siziman Bay	W	*Laricoxylon shilkinae* Blokhina	Oligocene	Blokhina (1984)
Primor'ye	V	*Larix primoriensis* Piemenov	Miocene	Piemenov (1986)
Botchi River	V	*Larix edelsteinii* Akhmetiev	Miocene	Akhmetiev (1973)
Aldan	C	*Larix omoloica* Dorofeev	Miocene	Dorofeev (1972)
Botchi River	C	*Larix schmidtiana* Akhmetiev	Miocene	Akhmetiev (1973)
Botchi River	W	*Larix* sp.	Miocene	Akhmetiev (1973)
Korf Bay	W	*Larix preobrajenskyi* Kryshtofovich	?Miocene	Kryshtofovich (1962)
Novy Log	W	*Larix* sp.	Oligocene	Dorofeev (1970)
Japan				
Hokkaido	V	*Larix* sp.	Pleistocene	Suzuki (1985)
Onbara	W	*Larix onbaraensis* Tanai and Onoe	Miocene	Tanai and Onoe (1961)
Poland				
Krościenko	V	*Larix ligulata* Szafer	Pliocene	Szafer (1946)

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[1] C = cones, P = pollen, V = vegetation (seeds, twigs, and needles), and W = wood.

Route for faunal migration has been recognized by McKenna (1972a) and West and Dawson (1978), who described a 30 per cent similarity between the mammalian fauna of North America and Europe during the Late Paleocene, rising sharply to over 50 per cent by the Middle Eocene, and dropping off to approximately 10 per cent following the Eocene. The rise and fall in the faunal similarities between Europe and North America reflect the establishment of this corridor in the early Tertiary, and its destruction in the Late Eocene/Early Oligocene (McKenna, 1972a; Dawson et al., 1975, 1976; West et al., 1977; West and Dawson, 1978; Hoch, 1983).

The Eocene-Oligocene boundary corresponds to a time when seafloor spreading in the Norwegian and Greenland seas split Svalbard off from northern Greenland, destroying the DeGeer Route (McKenna, 1972a; Talwani and Eldholm, 1977). The reconstruction in Figure 6 illustrates the separation of Svalbard and the broken land bridge in Late Eocene time.

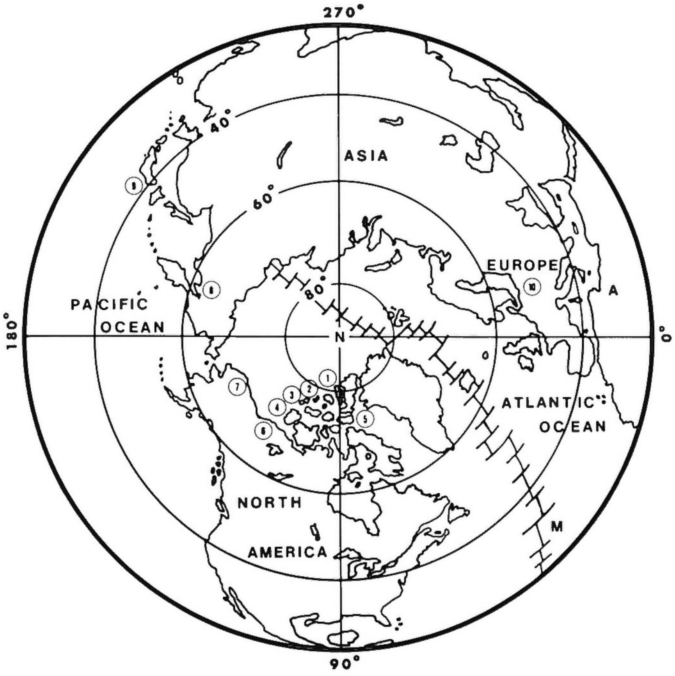

Figure 4. Map showing the present positions of continents in the Northern Hemisphere. Principal fossil localities containing the remains of ***Larix*** *have been numerically identified as: 1, Axel Heiberg Island; 2, Meighen Island; 3, Prince Patrick Island; 4, Banks Island; 5, Devon Island; 6, Yukon Territory; 7, Alaska; 8, Russia; 9, Japan; and 10, Poland. See Table 1 for specific details regarding the various species of* ***Larix*** *from these fossil sites. (Figures 4 to 7 are adapted from Florin, 1963; Hopkins, 1967; McKenna 1972a, 1975, 1983a; Raven and Axelrod, 1974; Hancock, 1975; Obradovich and Cobban, 1975; Williams and Stelck, 1975; Keigwin, 1978; Hancock and Kauffman, 1979; Smith et al., 1981; Hamilton, 1983; Irving, 1983; Nilsen, 1983; Thiede, 1983; Ziegler et al., 1983; Weimer, 1984; Tiffney, 1985a; Williams, 1986; Scotese and Denham, 1988; Scotese et al., 1988; and Ziegler, 1988.)*

The distribution of several Mesozoic fossil localities in North America and Europe that have some taxa of early conifers in common (Florin, 1963) suggests that the DeGeer Route was a land bridge for plant dispersal before Tertiary time. Although the evidence demonstrating that angiosperms used the DeGeer or Thulian routes is circumstantial, it would be unreasonable to assume that some angiosperm taxa did not use these corridors (Tiffney, 1985a). Raven and Axelrod (1974) suggested that the DeGeer Route (and perhaps the Thulian Route) may have been available well into the Neogene, although this would have involved long distance dispersal over open water.

The DeGeer Route apparently was intact and available to plants of the Canadian High Arctic during the time of deposition of the Buchanan Lake Formation. However, *Larix* and most other conifers of the Buchanan Lake fossil forests of Axel Heiberg Island are absent from the Paleocene and Eocene fossil record of Europe, as for example in southern England (see Chandler, 1961, 1962, 1963, 1964, 1978), and

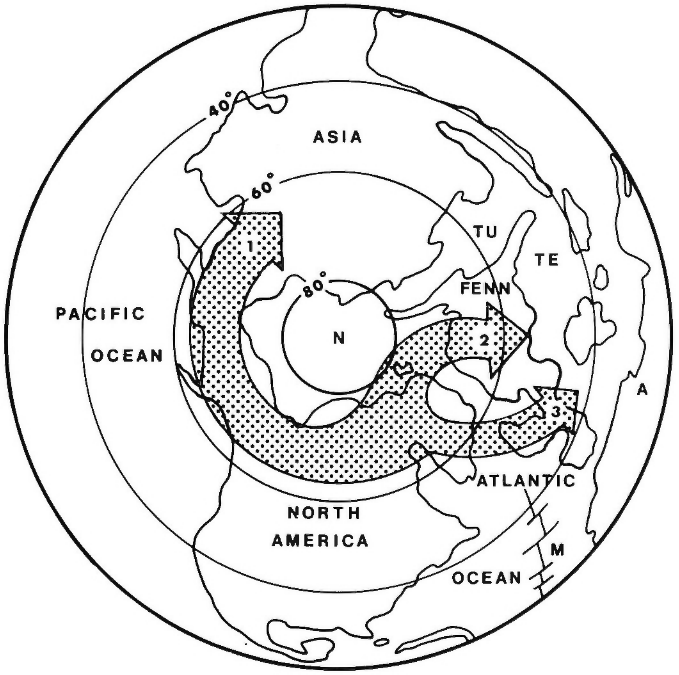

Figure 5. Generalized paleogeographic reconstruction of the Northern Hemisphere during the Paleocene (60 Ma), showing: 1, the Beringian Corridor; 2, the DeGeer Route; and 3, the Thulian Route. Legend: ***TU*** *= Turgai Strait,* ***TE*** *= Tethys Sea,* ***A*** *= Africa,* ***FENN*** *= Fennoscandia, and* ***M*** *= Mid-Atlantic Ridge.*

Larix does not appear in the fossil record of Europe until the Pliocene (Szafer, 1946). While the DeGeer Route seems to have been available for migration of taxa between North America and Fennoscandia-Europe, the warmer, tropical to subtropical climate of Europe during the Paleogene (Collinson, 1983) may have served as an effective filter. It seems likely that temperate taxa, such as those of the Buchanan Lake fossil flora, were thus precluded from utilizing this route into Europe during the Eocene, and that the destruction of the DeGeer land bridge at the end of the Eocene prevented later transatlantic dispersal when European climate was more temperate.

The Thulian Route

The Thulian Route, the southern route between Greenland and Europe, is thought to have existed along the Iceland-Faeroe Ridge during the Paleocene at a time of extension of rifting into the Norwegian-Greenland seas (McKenna, 1972a, 1975; Talwani et al., 1976; Berggren and Schnitker, 1983; Thiede and Eldholm, 1983). McKenna (1972b) argued that there are several problems associated with the Thulian Route, including the young age of the igneous rocks along this route and the depth below sea level to which certain sections of the Iceland-Faeroe Ridge have subsided. It is not certain that enough of the Thulian Route was emergent to provide an effective land route between North America and Europe. In any case, once deep water connections were established in Denmark Strait and the Iceland-Faeroe Channel in Late Eocene time, the intercontinental exchange of flora and fauna would have come to an end (Miller and Tucholke, 1983; Thiede and Eldholm, 1983). Also, the low paleolatitude of the Thulian Route would have hindered dispersal of temperate forms.

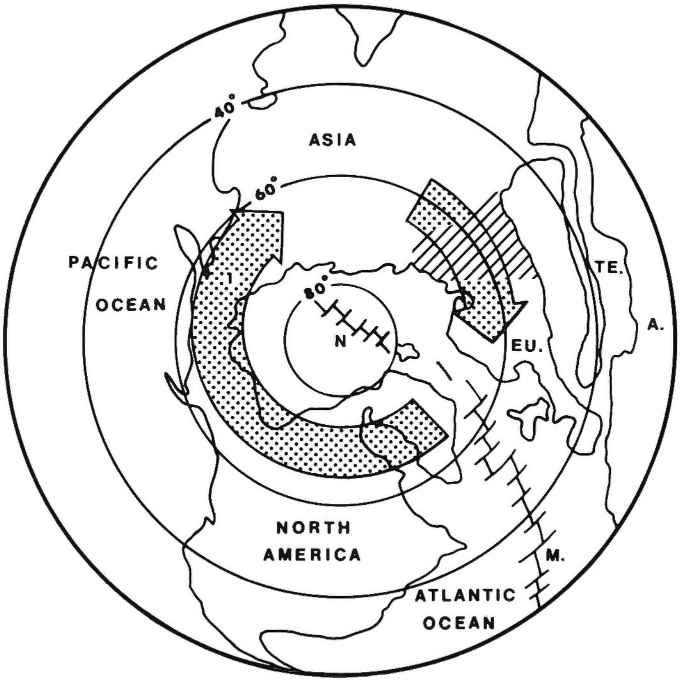

Figure 6. *Generalized paleogeographic reconstruction of the Northern Hemisphere for the Late Eocene (38 Ma), showing the Beringian Corridor (1). By the end of the Eocene, the DeGeer and Thulian routes between North America and Eurasia were broken. The stippled region indicates the location of the remnants of the Turgai Strait. Note the separation of Svalbard from Greenland. Legend:* ***TE*** *= Tethys Sea,* ***EU*** *= Europe,* ***A*** *= Africa, and* ***M*** *= Mid-Atlantic Ridge.*

The Beringian Corridor

Throughout much of Cretaceous and Tertiary time, Beringia was above sea level and was, therefore, a functional corridor for floral and faunal exchange between Asia and North America (Chaney, 1940, 1947; Hopkins, 1967; Wolfe, 1972, 1975; McKenna, 1983b; Tiffney, 1985a,b). Figure 4 illustrates the marked concentration of *Larix* fossil localities in the region of Beringia.

Beringia and the fossil forest locality on Axel Heiberg Island were at similar paleolatitudes during the Eocene (Fig. 7), and climate along this route may have been uniform. This is apparently confirmed by similarity of vegetation recovered from Alaska and the fossil forests (Chaney, 1947; Spicer et al., 1987; Basinger, *this volume*). Past and present distribution of short-bracted *Larix* species and the present distribution of long-bracted species indicate that the Beringian Route was used by *Larix* from Oligocene time and possibly earlier. It is not known whether the somewhat later appearance of *Larix* in Asia was a result of lack of preservation and discovery, or delayed dispersal from North America.

CONCLUSIONS

The genus *Larix* existed in the Canadian High Arctic in the Eocene and is known from many localities in northern North America and Asia in the Oligocene and Neogene. The distribution of *Larix* in the early Neogene parallels the spread of coniferous vegetation in the high northern latitudes. The high-latitude Beringian Corridor provided land connection between North America and Asia, forming a continuous northern landmass and facilitating dispersal.

The late Paleogene demise of the DeGeer and the Thulian dispersal routes of the North Atlantic Ocean, and the lower paleolatitude of these routes (and therefore warmer climate) apparently prevented the eastward dispersal of *Larix*, and perhaps many other northern temperate taxa, into Europe. The westward dispersal of *Larix* into Europe via northern Asia was blocked by the Turgai Strait, a shallow epicontinental seaway extending between the Arctic Ocean and Tethyan Sea throughout most of the Paleogene (Florin, 1963; McKenna, 1972a, 1975; Raven and Axelrod, 1974; Hamilton, 1983; Tiffney, 1985a). The Grand Coupure of Europe (McKenna, 1983b) shows a marked turnover in the composition of the western European fauna following the Eocene, when land connections with Asia were re-established. However, the fossil record of *Larix* in Europe indicates that this genus did not become established in Europe until the Pliocene. The apparent affinity between the extant European *L. decidua* and *L. decidua* var. *polonica* and the Asian *L. sibirica* provides further evidence of relatively recent dispersal of the genus into Europe from Asia. Pre-Pliocene entry of *Larix*, and perhaps other boreal taxa, into Europe may have been limited by climatic factors. The Tertiary distribution of *Larix* and the known paleogeographic history of the Northern Hemisphere indicate that transatlantic dispersal of northern temperate taxa, particularly boreal coniferous forms, was improbable, or at best restricted to a few species.

Review of the distribution of fossil and living members of the genus indicates that short-bracted species of *Larix* have been prominent members of the high-latitude boreal forests throughout much of the middle and late Cenozoic. The long-bracted larches, so far not known or absent from the fossil record, are restricted in range, are often found in isolated populations, are confined to montane habitats, and have an amphi-Pacific distribution. The long-bracted larches have probably been distinct from the short-bracted forms since the Paleogene, and based on the amphi-Pacific distribution pattern of the modern long-bracted forms, early Tertiary representatives of this group apparently followed two Tertiary routes of dispersal from a high-latitude source: the western North American and the eastern Asian (Fig. 7).

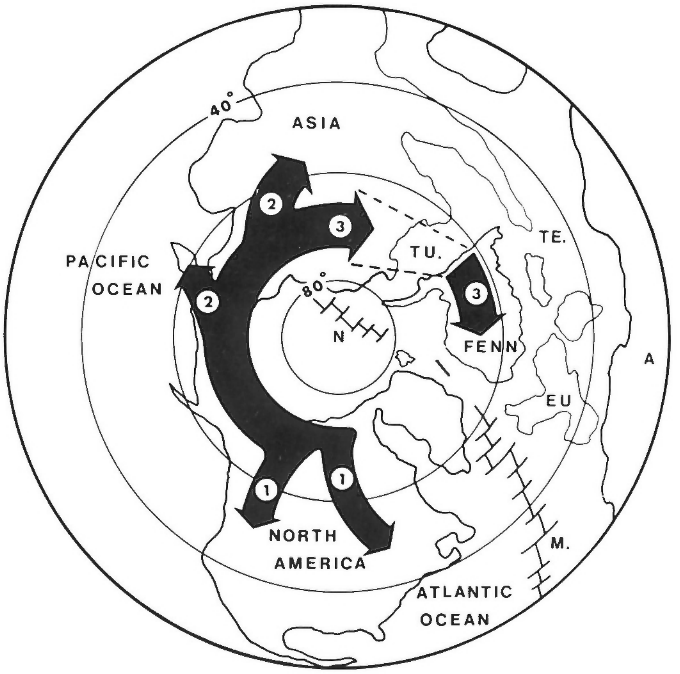

Figure 7. *Generalized paleogeographic reconstruction of the Northern Hemisphere during the Middle Eocene (45 Ma), showing the possible dispersal routes of* ***Larix*** *into Canada, Asia, and Western Europe. Dispersal into northern Europe was delayed until the Pliocene. Legend: 1, the North American Route; 2, the Southern Asian Route; and 3, the Trans-Eurasian Route.* ***TU*** *= Turgai Strait,* ***TE*** *= Tethys Sea,* ***A*** *= Africa,* ***FENN*** *= Fennoscandia,* ***EU*** *= Europe, and* ***M*** *= Mid-Atlantic Ridge.*

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POLLEN AND SPORE FLORA OF AN EOCENE FOREST, EASTERN AXEL HEIBERG ISLAND, N.W.T.

D.J. McIntyre[1]

McIntyre, D.J., 1991. Pollen and spore flora of an Eocene forest, eastern Axel Heiberg Island, N.W.T. In Tertiary Fossil Forests of the Geodetic Hills, Axel Heiberg Island, Arctic Archipelago, R.L. Christie and N.J. McMillan (eds.); Geological Survey of Canada, Bulletin 403, p. 83-97.

Abstract

The fossil forest succession east of Geodetic Hills, Axel Heiberg Island, occurs within the Buchanan Lake Formation, Eureka Sound Group. The distribution of dominant and abundant pollen and spore taxa in the different horizons shows that there were numerous floral changes during the period of deposition and that forest was not always the dominant vegetation type. These floral changes, some possibly indicating successional vegetation types, suggest the occurrence of a mosaic of plant communities in a continually changing floodplain environment. The pollen assemblages indicate a temperate to warm temperate, moist climate, with only slight suggestion of subtropical or tropical influence. The rich and diverse pollen flora are similar to Middle Eocene assemblages from other parts of the Canadian Arctic. They are probably Middle Eocene in age, but a Late Eocene age cannot be discounted. Significant pollen taxa in the Geodetic Hills assemblages include *Metasequoia* sp., *Tsuga* sp., *Anacolosidites* sp., *Carya* spp., *Engelhardtia* sp., *Gothanipollis* sp., *Intratriporopollenites* sp., *Juglans* sp., *Pistillipollenites mcgregorii*, and *Tilia* sp.

Résumé

La succession de forêts fossiles, située à l'est des collines Geodetic dans l'île Axel Heiberg, fait partie de la formation de Buchanan Lake, du groupe d'Eureka Sound. La distribution des taxons abondants et dominants, représentés par des pollens et par des spores, dans les différents horizons, montre qu'il s'est produit de nombreuses variations floristiques durant la période de sédimentation, et que la forêt n'était pas toujours le type végétal dominant. Ces variations floristiques, dont quelques-unes indiquent peut-être une succession des types de végétation, suggèrent la présence d'une mosaïque de communautés végétales dans un milieu de plaine inondable subissant un changement continuel. Les assemblages de pollens indiquent une température tempérée à chaude, un climat humide, et seulement une légère possibilité d'influences subtropicales ou tropicales. La flore pollinique riche et diversifiée ressemble aux assemblages de l'Éocène moyen d'autres régions de l'Archipel arctique canadien. Elle date probablement de l'Éocène moyen, mais un âge correspondant à l'Éocène supérieur est quand même possible. Parmi les taxons polliniques significatifs contenus dans les assemblages des collines Geodetic, citons *Metasequoia* sp., *Tsuga* sp., *Anacolosidites* sp., *Carya* spp., *Engelhardtia* sp., *Gothanipollis* sp., *Intratriporopollenites* sp., *Juglans* sp., *Pistillipollenites mcgregorii* et *Tilia* sp.

INTRODUCTION

Well preserved plant material, particularly stumps, logs and woody and leafy litter, occurs in intervals one to two metres thick in the Buchanan Lake Formation, at a site 25 km east of Geodetic Hills, Axel Heiberg Island. These intervals, especially those that contain numerous in situ stumps, form the Geodetic Hills fossil forest. In addition to the stumps and wood of the fossil forest (see Francis, *this volume*) and the leaf, cone, and seed material (see Basinger, *this volume*), excellent pollen floras have been recovered from many horizons in the forest interval, including the leaf litter layers and the intervening silts and mudstones.

In 1987, seventy samples (GSC locs. C-153316-350, C-163151-185) were collected at varied intervals from all lithological types from a section (MDB-87-1) 238 m thick at the fossil forest locality (Fig. 1). The forest layers (approximately 80 m of section from 60 to 140 m) and strata both below and above the forest interval were sampled. The section lies 400 m east of the main area of fossil stumps at lat. 79°55'N, long. 88°58'W (NTS 59 H). The sediments consist of interbedded sandstones, grey to occasionally black siltstones and mudstones, and black organic layers (up to 1 m thick) of leafy and woody detritus. Logs and in situ stumps are present in many of the strata, especially in the middle part of the section.

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[1] Geological Survey of Canada, Institute of Sedimentary and Petroleum Geology, 3303 - 33rd Street N.W., Calgary, Alberta T2L 2A7

The strata of the fossil forest area belong to the Buchanan Lake Formation, Eureka Sound Group (Ricketts, 1986) and are considered to be at the top of the formation (Ricketts and McIntyre, 1986). Preliminary palynological results from a few samples collected close to the section of the present study indicated a Middle Eocene age for this interval (Ricketts and McIntyre, 1986). Bustin (1982) reported a Miocene/(?)Pliocene age for strata east of Geodetic Hills, based on microfloral determinations, but these assemblages were recently shown to be Middle Eocene in age (Ricketts and McIntyre, 1986). These strata are, therefore, considered to be coeval with the fossil forest strata and thus belong to the Buchanan Lake Formation.

POLLEN AND SPORE ASSEMBLAGES

Most of the samples processed yield adequate pollen and spore assemblages in which the dominant types are usually Taxodiaceae (23 samples) (most of which are probably *Metasequoia*), *Alnus* (14 samples), and Pinaceae (13 samples) (mainly *Picea* with rare *Pinus* and possibly very rare *Abies*). Other pollen and spore types that are occasionally dominant are *Betula* (3 samples), undifferentiated tricolpate angiosperm pollen (3 samples), Ericaceae and *Laevigatosporites* (one sample each). In almost all samples, one or more of the groups or genera discussed above is the second most abundant form. Pollen flora composition is generally similar throughout the section, although dominance and abundance of the taxa vary from sample to sample. Most of the pollen and spore taxa recorded are common, and sometimes abundant, in the rich palynological preparations. However, except for the dominant forms mentioned above, any one taxon rarely constitutes more than five per cent of the total assemblage. Some of the taxa that appear to be significant for floral, climatic and environmental interpretation are discussed below.

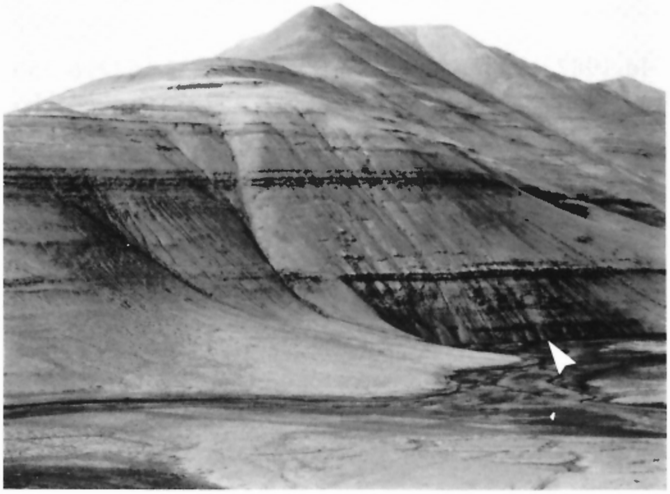

Figure 1. *Section MDB-87-1, fossil forest succession. Arrow indicates base of section.*

The taxa identified in the fossil forest section are listed below. Most of them are illustrated in Plates 1 to 4.

Pteridophyta

Lycopodium spp. (? *L. annotinum* L., *L. inundatum* L., *L. obscurum* L.)
Gleichenia sp.
Osmunda spp.
?Polypodium sp.
?Cryptogramma sp.
Laevigatosporites spp.
Deltoidospora spp.
Radialisporis radiatus (Krutzsch) Krutzsch

Spermatophyta

Gymnospermae, Coniferae
?Abies sp.
?Larix sp.
Picea spp.
Pinus sp.
Tsuga sp.
?Cupressaceae
Taxodiaceae (mainly *Metasequoia* sp., probable *Glyptostrobus* sp., possible *Taxodium* sp.)

Angiospermae, Monocotyledonae
Liliacidites sp.
Monocolpopollenites sp.
Sparganium sp.

Angiospermae, Dicotyledonae
Acer sp.
Alnus spp.
Anacolosidites sp. cf. *A. reklawensis* Elsik
Betula spp.
Carya spp. including *C. veripites* Wilson and Webster, *C. viridifluminipites* Wodehouse and *C.* sp. (undescribed)
Castanea sp.
Cercidiphyllum sp.
Corylus sp.
Diervilla sp.
Engelhardtia sp. cf. *E. chrysolepis*
Ericaceae (*Ericipites* spp.)
Fagus sp.
Fraxinus sp.
Gothanipollis sp.
Ilex sp.
Intratriporopollenites sp. (cf. *Reevesia*)
Juglans spp.
Liquidambar sp.
Lonicera sp.
Myrica sp.
Nyssa sp.
Pachysandra sp.
Pistillipollenites mcgregorii Rouse
?Planera sp.

Pterocarya spp.
Quercus spp.
?Rhus sp.
Rosaceae
Salix sp.
Tilia sp. (*T. vescipites* Wodehouse)
Tricolpate pollen, undifferentiated, many species
Tricolporate pollen, undifferentiated, many species
Tricolporopollenites kruschii (Potonie) Thomson and Pflug
Tricolporopollenites sp. A
Ulmus sp. (*Ulmipollenites undulosus* Wolff)
?Viburnum sp. cf. *V. cassinoides* L.

Some taxa that occur consistently through the section in relatively small numbers may appear to be common to abundant in a few samples. However, pollen counts show that these taxa usually constitute less than one per cent of the assemblage of any sample. Such taxa include *Carya*, *Tilia*, *Quercus*, *Corylus*, *Intratriporopollenites*, *Ulmus*, *Juglans*, *Pterocarya*, *?Viburnum*, *Liliacidites* and *Engelhardtia*. In a few samples, however, *Carya* (78 m, 86.4 m), *Pterocarya* (67.8 m, 141 m), *Ulmus* (3 m) or *?Viburnum* (142 m) may constitute up to five per cent of the total pollen and spores. Pollen of *Engelhardtia* (59 m, 62.5 m, 65.6 m, 86.4 m) and *Liliacidites* (26.5 m, 27.5 m, 67.8 m) may constitute up to 13 per cent of the pollen content in a few samples. Taxa that occur sporadically throughout the section but are never common include *Acer*, *Castanea*, *Cercidiphyllum*, *Fagus*, *Fraxinus*, *Gothanipollis*, *Ilex*, *Liquidambar*, *Pachysandra*, *Sparganium*, *?Larix*, *Tricolporopollenites* sp. A and *T. kruschii*. Many of the undifferentiated tricolpate and tricolporate pollen types also belong to this group of occasionally occurring taxa. Certain pollen types that occur sporadically in the section are common or abundant in a few samples. These include *Polypodium*, *Diervilla*, *Anacolosidites*, and *Pistillipollenites mcgregorii*.

Although there is variation from sample to sample in the dominance and abundance of species and in the presence or absence of particular species, no obvious pattern of palynofloral change is discernible and the pollen floras are generally similar throughout. Neither is there any obviously consistent difference in the pollen floras from the different lithotypes, whether it be diversity, dominance, or abundance of species present. The differences between the pollen and spore assemblages in the section are considered here to be of little biostratigraphic significance. They represent parts of a regional flora and apparently belong to one biostratigraphic unit.

VEGETATION

The dominance and abundance of pollen of Taxodiaceae (apparently mainly *Metasequoia*) and Pinaceae (mainly *Picea*) indicate that, for much of the time represented by the pollen floras of the fossil forest section, forests of *Metasequoia* and *Picea* were significant elements of the vegetation of the Geodetic Hills region, but perhaps not always immediately adjacent to the site of deposition. The palynological evidence suggests that forest was often not the vegetation type near sites of deposition. The dominance of *Alnus* pollen in some samples, and its abundance in many others, suggests that forest margin or shrub vegetation commonly occurred close to, or at, the sites of deposition. The occasional dominance of fern spores (*Laevigatosporites*) or Ericaceae pollen suggests that areas of open wetland scrub were at times dominant, at least locally. In other intervals in the sequence, the pollen types present suggest that there was mainly woodland of mixed angiosperms, possibly small trees and shrubs. The abundance, in numerous samples, of pollen such as that of *Betula*, *Carya*, *Corylus*, *Engelhardtia*, *Juglans*, *Liquidambar*, *Pterocarya*, *Quercus*, *Tilia* and *Ulmus*, as well as a variety of other angiosperms, provides strong evidence that mixed woodland and/or shrubland was significant in the area.

The palynofloras indicate that varied and different plant communities occupied the area during the time of deposition of the fossil forest succession. It is probable that most of the palynofloras represent local vegetation types and the true regional flora is not properly reflected in any one sample. There were undoubtedly numerous and continuous successional vegetational changes and, for most of the time (perhaps the entire time) a mosaic of plant communities existed in the region, responding to a continually changing environment. The pollen flora of a particular sample may, therefore, represent only a very limited part of this mosaic. It is noteworthy that none of the different palynological assemblages representing this mosaic is confined to one specific lithological type.

ENVIRONMENT

The mosaic of plant communities represented by the various pollen floras is undoubtedly a product of the constantly changing local environment during deposition of the fossil forest succession. The pollen floras suggest that many of the vegetation changes in the fossil forest interval were of a successional nature in a continually changing floodplain environment. The pollen assemblages also indicate (McIntyre, work in progress) that there is lateral variation along at least some of the organic horizons.

The muds and silts appear to be overbank deposits, whereas the organic layers, both leaf mat and woody, were deposited in swamps (Ricketts, *this volume*). The association of stumps with such horizons provides further evidence of swamp conditions, which apparently were often quite extensive, considering the continuity of these layers over distances exceeding 2 km. The swamp environment was probably occupied most commonly by a climax forest in which *Metasequoia* was the dominant tree. Nearby drier upland areas likely provided the habitat for *Picea*-dominant forests. The mixed angiosperm forest and shrub associations could have grown on low-lying areas that were possibly subject to more frequent or prolonged flooding. The occasional abundance of fern spores and *Liliacidites* pollen suggests that open areas, mostly free of shrub and tree cover, were periodically present. The presence of river and stream margin environments, perhaps sites of early successional stages preceding forest development, is suggested by the dominance and abundance of *Alnus* in many samples.

CLIMATE

Most of the pollen and spore types recorded from the fossil forest strata are from genera that today grow in temperate to warm temperate regions. The presence of pollen of *Carya*, *Corylus*, *Acer*, *Juglans*, *Tilia*, *Ulmus*, *Diervilla*, *Viburnum*, *Fraxinus*, *Pterocarya*, *Alnus*, *Fagus*, *Quercus*, *Lonicera*, *Larix*, *Picea* and *Tsuga* in association suggests a temperate to warm temperate climate. These genera occur today in parts of eastern North America and eastern Asia, at latitudes of 30° to 50°, where temperate to warm temperate climates prevail (Axelrod, 1984; Hou, 1983; Wolfe, 1979). Such areas have limited spells of cool winter weather in which temperatures a few degrees below freezing commonly occur, and less frequent short spells of more severe winter conditions to –15°C are known. Therefore, it seems reasonable to assume that the fossil forest vegetation also experienced cool temperatures.

The presence of *Metasequoia*, *Glyptostrobus* and *Engelhardtia* further suggests conditions that were more warm temperate than temperate. *Anacolosidites* is the only pollen type that suggests a subtropical or tropical influence, but its presence may simply imply a wider climatic tolerance in the Tertiary for the parent plants. The *Anacolosidites* pollen species present appears to be more similar morphologically to those of the extant South American genus *Cathedra* than to those of the extant Indomalaysian and Pacific genus *Anacolosa*. Both genera are members of the tropical family Olacaceae. The evidence from the fossil forest palynological assemblages strongly suggests that coniferous forests (both evergreen and deciduous) and deciduous broadleaf angiosperm hardwood forests flourished in the Axel Heiberg area of the Arctic, in the early Tertiary, in a temperate to warm temperate climate. The present latitude (80°N) is considered (McKenna, 1980; Irving and Wynne, *this volume*) to be only a few degrees higher (at most) than the paleolatitude at the time the forests flourished. For such forests to prosper in the warm temperatures postulated, a significant amount of moisture would be necessary. It is likely that the rainfall in the area was high, and similar (in the range of 1000 to 1500 mm annually) to that recorded for analogous forests of the present day. The paucity of adequate light at these high latitudes during winter may be of little significance because of the deciduous nature of most of the plants (see Basinger, *this volume*).

AGE

The strata of the fossil forest succession belong to the Buchanan Lake Formation (Ricketts, *this volume*), which, from the occurrence of palynofloras in a few sections, has been determined as Middle Eocene in age (Ricketts and McIntyre, 1986). The fossil forest macrofloras have also been determined to be of Eocene age (Basinger, 1987; *this volume*). Vertebrate faunas from the Iceberg Bay Formation, immediately below the Buchanan Lake Formation stratigraphically, have been determined to be of Early or early Middle Eocene age (West et al., 1977; McKenna, 1980; Estes and Hutchison, 1980). The palynofloras of the fossil forest section are stratigraphically above the youngest pollen assemblages discussed by Choi (1983) and by McIntyre (*in* Ricketts, in press), which were considered as Middle Eocene by Choi and Early to (?)Middle Eocene by McIntyre. The youngest pollen floras of Choi and McIntyre both contain *Novemprojectus traversii* Choi, which is absent from the fossil forest samples. However, *Juglans*, *Tsuga* and *Engelhardtia*, which are commonly present in the fossil forest samples, are not present in the youngest palynofloras described by either Choi (1983) or McIntyre (*in* Ricketts, in press). Pollen floras of the fossil forest succession are similar to those recorded from the Middle and Late Eocene of the Arctic by Rouse (1977). The Oligocene assemblages documented by Piel (1971), Rouse (1977) and Ioannides and McIntyre (1980) are generally similar to those from the fossil forest succession. However, some distinctive species recorded from the Oligocene are lacking in the Geodetic Hills assemblages.

Many of the pollen taxa present in the fossil forest strata are important members of Eocene assemblages in the Arctic and farther south, but most are not restricted to the Eocene. *Carya veripites* and *C. viridifluminipites*, both common in the Eocene and Oligocene of the Arctic, are often abundant in the fossil forest section. *Tilia vescipites*, usually common in the fossil forest section, is generally common in Eocene strata of the Arctic, but is extremely rare in the Paleocene (McIntyre, 1989, McIntyre *in* Ricketts, in press). It was not recorded in the Oligocene of the Arctic by Rouse (1977) but occurs in Oligocene strata in other parts of Canada (Piel, 1971; Ioannides and McIntyre, 1980; Norris, 1986).

Pollen of *Tsuga* occurs consistently in the fossil forest succession and is more common in the upper part of the section. It was recorded in Lower or Middle Eocene strata by Ioannides and McIntyre (1980), but neither Choi (1983) nor McIntyre (*in* Ricketts, in press) have noted this pollen type in the Eocene of other sections of the Eureka Sound Group. *Juglans* was not recorded by either Choi (1983) or McIntyre (*in* Ricketts, in press) from the Early or (?)Middle Eocene. It apparently first appears in strata of Early or Middle Eocene age in the Mackenzie Delta (Ioannides and McIntyre, 1980) and of Early-Middle Eocene age in the Arctic (Rouse, 1977) and is common to abundant in many samples of the fossil forest succession.

Engelhardtia, a common and sometimes abundant member of the fossil forest palynofloras, was recorded by Rouse (1977) as *Momipites coryloides* Form A, from strata of Early-Middle Eocene age in the Arctic. *Intratriporopollenites*, which occurs sporadically in the fossil forest succession, was also recorded from Lower to Middle Eocene strata by Rouse (1977). Both of these species appear to have restricted ranges in the Arctic and have not been recorded there in the Paleocene.

Tricolporopollenites kruschii, a rather rare species in this section, was recorded from the Early-Middle Eocene by Rouse (1977) and from the Eocene by Doerenkamp et al. (1976). The Early-Middle Eocene *Lonicera* (Rouse, 1977) occurs rarely. Pollen of *Diervilla*, not known before the Eocene, occurs sporadically throughout the section and is abundant in a few intervals. *Gothanipollis* occurs extremely rarely in the fossil forest interval. Rouse (1977) recorded it from the Upper Eocene and Elsik and Dilcher (1974) discussed its importance for dating of Eocene, especially Middle Eocene, strata in North America.

The *Anacolosidites* recorded in the fossil forest succession appears to be very close to *A. reklawensis* described from the Middle Eocene of the Gulf Coast, U.S.A., by Elsik (1974). The presence of *Pistillipollenites mcgregorii* in some samples in the lower half of the section provides important evidence for an age not younger than Eocene. This species is generally considered not to range above the Middle Eocene (Rouse, 1977), but Elsik (1974) indicates that it may occur as high as the base of the Upper Eocene.

Although many of the pollen species recorded from the fossil forest section range through the Eocene into younger strata in the Arctic and elsewhere, the available evidence indicates that the succession examined is not older than Middle Eocene. It is stratigraphically above strata determined to be Lower to (?)Middle Eocene (McIntyre, *in* Ricketts, in press) and it contains numerous palynomorphs that most likely first appear in the Middle Eocene. No significant change of age is apparent through the section, and it is either still Middle Eocene at the top or possibly early Late Eocene. There is no evidence to indicate that any part of the section is as young as Oligocene.

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PLATES 1 TO 4

Slides containing the figured specimens are curated in the type collection of the Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario K1A 0E8. They are at present in temporary storage at the Institute of Sedimentary and Petroleum Geology, Calgary, Alberta, where all duplicate slides are permanently stored. In the descriptions for Plates 1 to 4, the species name is followed by the GSC locality number (prefixed C), the slide number (prefixed P) and the GSC type number (prefixed GSC). Stage coordinates and England Finder readings for Reichert-Jung Polyvar microscope 392166 at the Institute of Sedimentary and Petroleum Geology, Geological Survey of Canada, Calgary, Alberta are on file with the curated specimens. All figures in the four plates are shown at a magnification of x1000.

PLATE 1

Figure 1. *Lycopodium annotinum*
C-163170, P3091-55f, GSC 96288.

Figure 2. *Lycopodium annotinum*
C-153319, P3091-4b, GSC 96289.

Figure 3. *Lycopodium inundatum*
C-163160, P3091-45f, GSC 96290.

Figure 4. *Lycopodium selago*
C-153323, P3091-8b, GSC 96291.

Figure 5. *Sphagnum* sp.
C-163170, P3091-55g, GSC 96292.

Figure 6. *?Polypodium* sp.
C-153337, P3091-22b, GSC 96293.

Figure 7. *?Polypodium* sp.
C-153337, P3091-22a, GSC 96294.

Figure 8. *?Cryptogramma* sp.
C-153323, P3091-8b, GSC 96295.

Figure 9. *Radialisporis radiatus*
C-153331, P3091-16i, GSC 96296.

Figure 10. *Deltoidospora* sp.
C-153331, P3091-16i, GSC 96297.

Figure 11. *Osmunda* sp.
C-163163, P3091-48b, GSC 96298.

Figure 12. *Osmunda* sp.
C-153339, P3091-24b, GSC 96299.

Figure 13. *Laevigatosporites* sp.
C-153331, P3091-16i, GSC 96300.

Figure 14. *Glyptostrobus* sp.
C-163159, P3091-44g, GSC 96301.

Figure 15. *Metasequoia* sp.
C-163160, P3091-45e, GSC 96302.

Figure 16. *Metasequoia* sp.
C-163151, P3091-36b, GSC 96303.

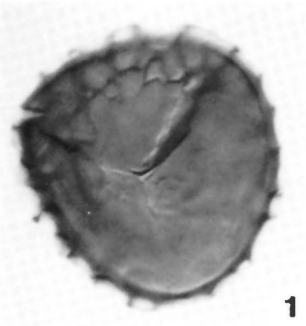
Figure 17. *Glyptostrobus* sp.
C-163178, P3091-63g, GSC 96304.

Figure 18. *?Larix* sp.
C-163165, P3091-50f, GSC 96305.

Figure 19. *Metasequoia* sp.
C-153342, P3091-27b, GSC 96306.

Figure 20. *Metasequoia* sp.
C-153342, P3091-27j, GSC 96307.

Figure 21. *Tsuga* sp.
C-163178, P3091-63h, GSC 96308.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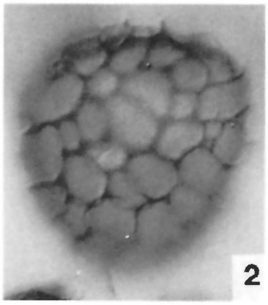
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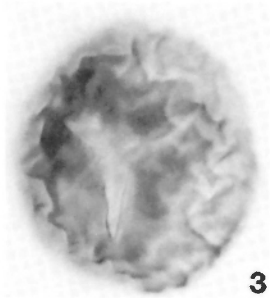
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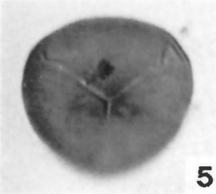
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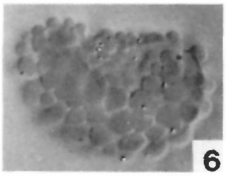
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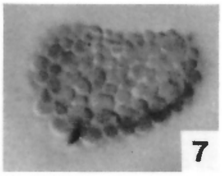
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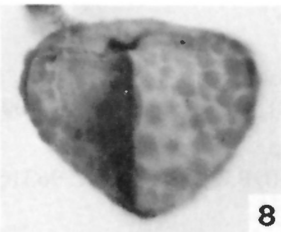
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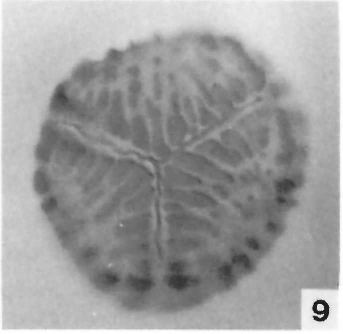
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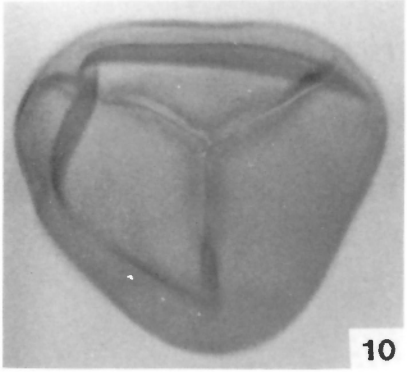
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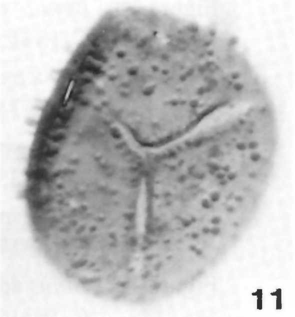
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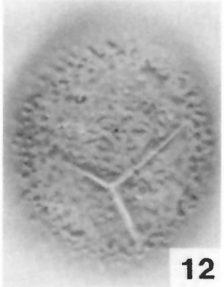
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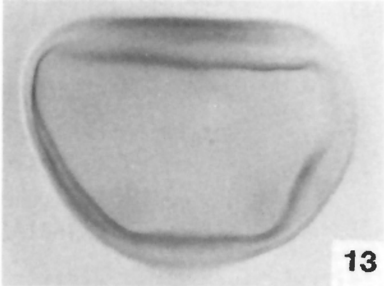
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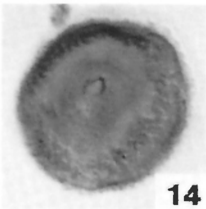
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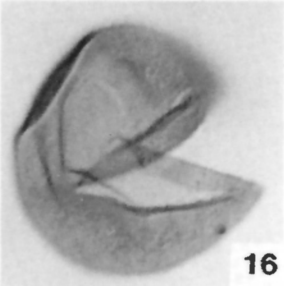
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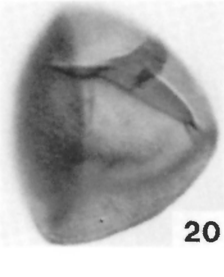
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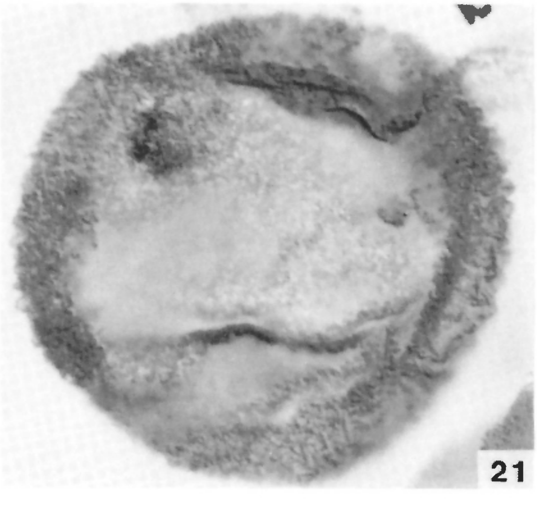
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PLATE 2

Figure 1. *Tsuga* sp.
C-153331, P3091-16i, GSC 96309.

Figure 2. *?Abies* sp.
C-163160, P3091-45e, GSC 96310.

Figure 3. *Picea* sp.
C-163159, P3091-44f, GSC 96311.

Figure 4. *Pinus* sp.
C-163151, P3091-36b, GSC 96312.

Figure 5. *Picea* sp.
C-163160, P3091-45f, GSC 96313.

Figure 6. Ericaceae
C-163170, P3091-55f, GSC 96314.

Figure 7. Ericaceae
C-163155, P3091-40b, GSC 96315.

Figure 8. *Alnus* sp.
C-163173, P3091-58b, GSC 96316.

Figure 9. *Corylus* sp.
C-163173, P3091-58c, GSC 96317.

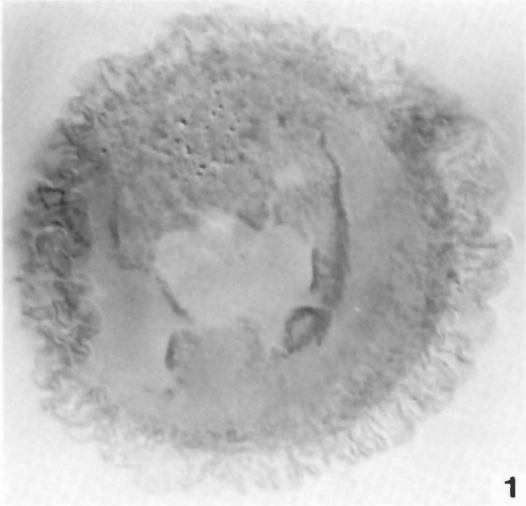
Figure 10. *Betula* sp.
C-163151, P3091-36c, GSC 96318.

Figure 11. *Myrica* sp.
C-153342, P3091-27h, GSC 96319.

Figure 12. *Alnus* sp.
C-153331, P3091-16j, GSC 96320.

Figure 13. *Alnus* sp.
C-163179, P3091-64c, GSC 96321.

Figure 14. *Alnus* sp.
C-163161, P3091-46c, GSC 96322.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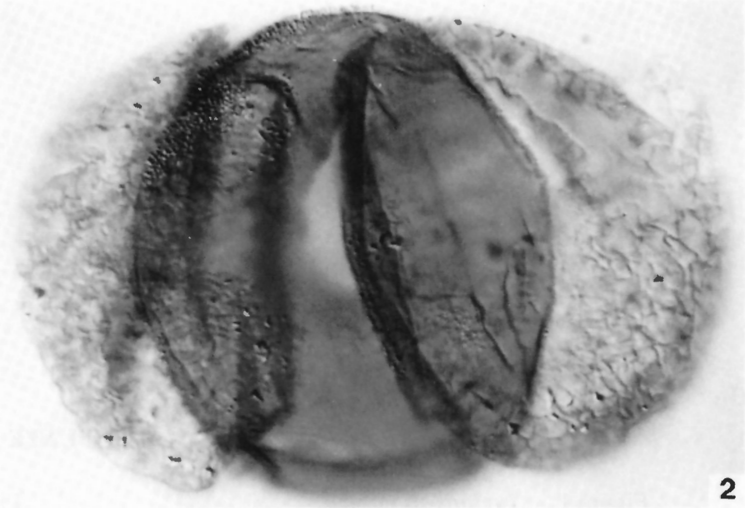
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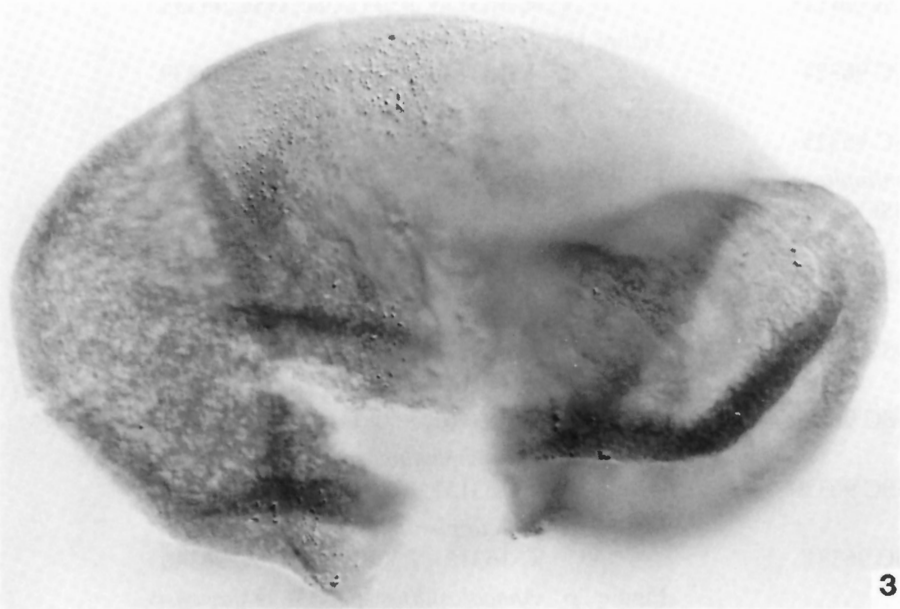
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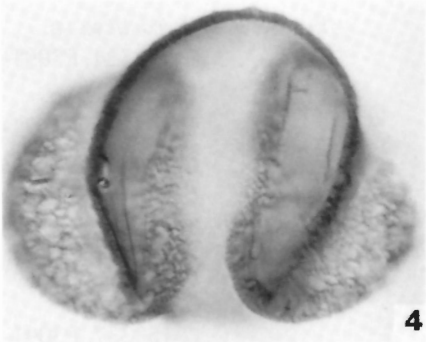
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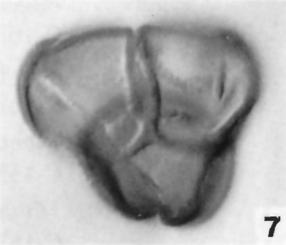
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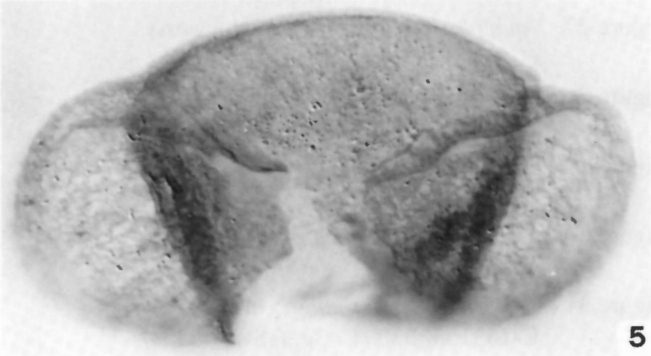
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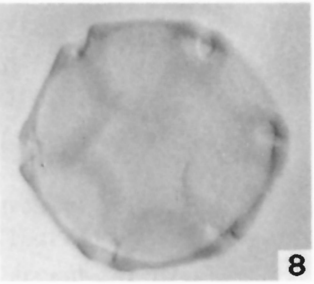
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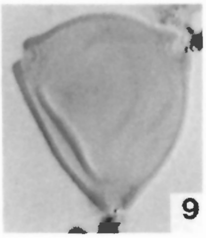
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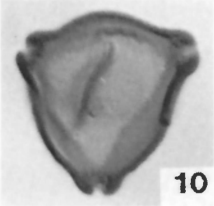
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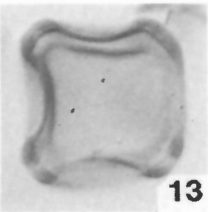
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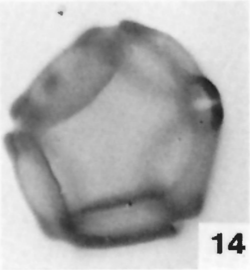
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PLATE 3

Figure 1. *Juglans* sp.
C-163170, P3091-55f, GSC 96323.

Figure 2. *Pterocarya* sp.
C-153324, P3091-9c, GSC 96324.

Figure 3. *Pterocarya* sp.
C-153324, P3091-9c, GSC 96325.

Figure 4. *Carya* sp. cf. *C. viridifluminipites*
C-163173, P3091-58b, GSC 96326.

Figure 5. *Carya veripites*
C-163151, P3091-36b, GSC 96327.

Figure 6. *Carya* sp.
C-153337, P3091-22c, GSC 96328.

Figure 7. *Engelhardtia* sp.
C-163151, P3091-36c, GSC 96329.

Figure 8. *Engelhardtia* sp.
C-163151, P3091-36c, GSC 96330.

Figure 9. *Ulmus* sp.
C-153324, P3091-9c, GSC 96331.

Figure 10. *?Planera* sp.
C-163151, P3091-36b, GSC 96332.

Figure 11. *Tilia* sp.
C-153339, P3091-24b, GSC 96333.

Figures 12, 13. *Tilia* sp.
C-163151, P3091-36b, GSC96334.

Figure 14. *Intratriporopollenites* sp.
C-163173, P3091-58c, GSC 96335.

Figure 15. *Intratriporopollenites* sp.
C-163173, P3091-58c, GSC 96336.

Figure 16. *Ilex* sp.
C-163175, P3091-60d, GSC 96337.

Figure 17. *Ilex* sp.
C-163173, P3091-58c, GSC 96338.

Figure 18. *Liquidambar* sp.
C-163155, P3091-40b, GSC 96339.

Figure 19. *?Viburnum* sp. cf. *V. cassinoides*
C-153324, P3091-9c, GSC 96340.

Figure 20. *?Viburnum* sp. cf. *V. cassinoides*
C-163151, P3091-36b, GSC 96341.

Figure 21. *?Viburnum* sp.
C-163151, P3091-36b, GSC 96342.

Figure 22. *Gothanipollis* sp.
C-153328, P3091-13j, GSC 96343.

Figure 23. *Tricolporopollenites* sp. A
C-163161, P3091-46b, GSC 96344.

Figure 24. *Cercidiphyllum* sp.
C-163151, P3091-36b, GSC 96345.

Figure 25. *Cercidiphyllum* sp.
C-163151, P3091-36b, GSC 96346.

Figure 26. *Anacolosidites* sp. cf. *A. reklawensis*
C-163151, P3091-36c, GSC 96347.

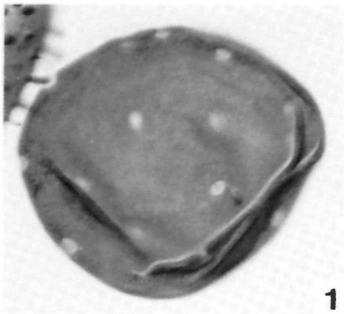
Figure 27. *Anacolosidites* sp. cf. *A. reklawensis*
C-163151, P3091-36c, GSC 96348.

Figure 28. *Pistillipollenites mcgregorii*
C-163151, P3091-36c, GSC 96349.

Figure 29. *Pistillipollenites mcgregorii*
C-163151, P3091-36c, GSC 96350.

Figure 30. *Pistillipollenites mcgregorii*
C-163151, P3091-36c, GSC 96351.

Figure 31. *?Pistillipollenites mcgregorii*
C-163158, P3091-43g, GSC 96352.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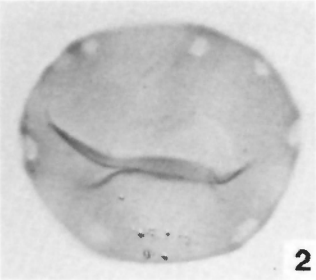
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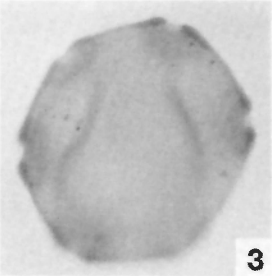
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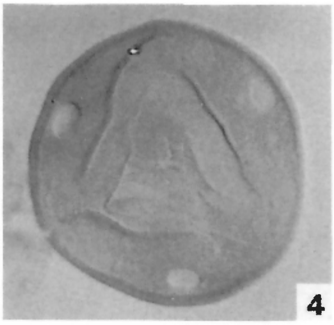
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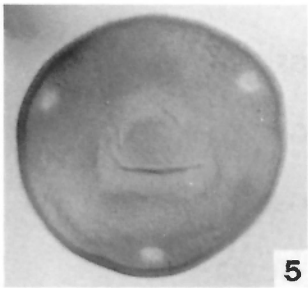
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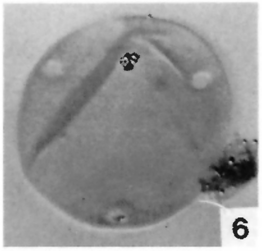
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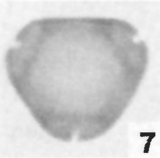
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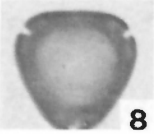
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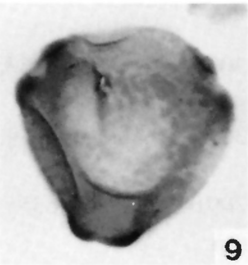
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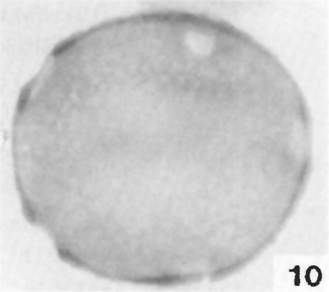
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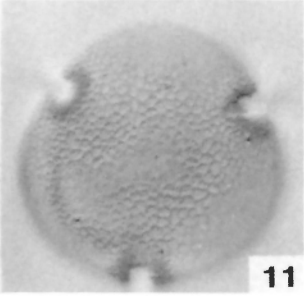
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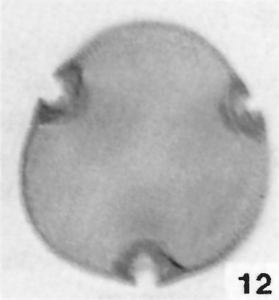
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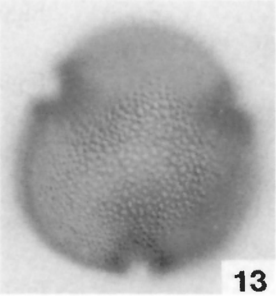
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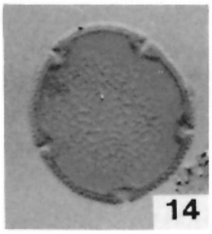
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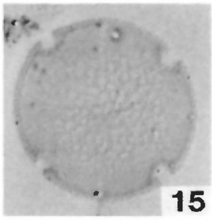
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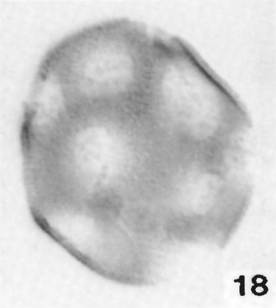
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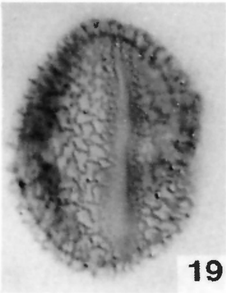
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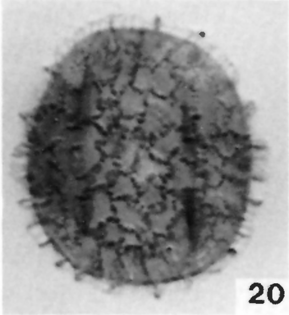
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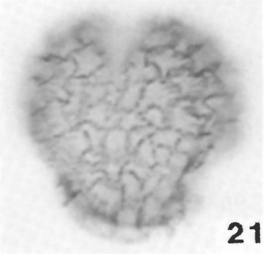
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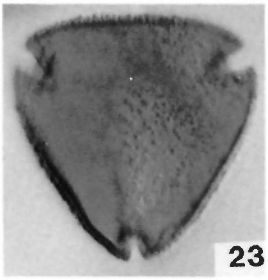
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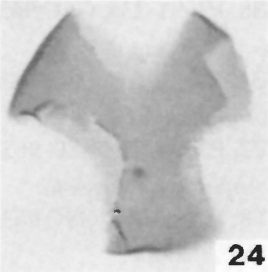
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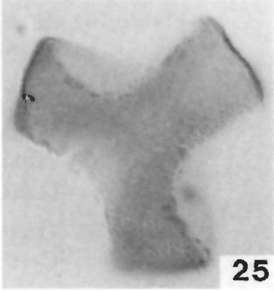
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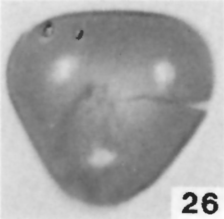
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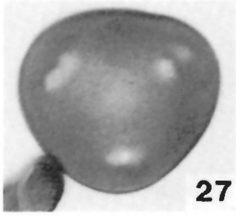
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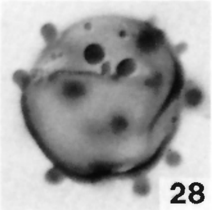
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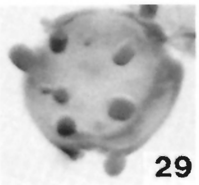
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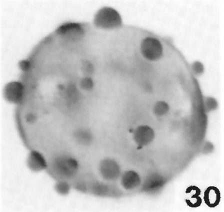
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PLATE 4

Figure 1. *Diervilla* sp.
C-163163, P3091-48b, GSC 96353.

Figure 2. *Diervilla* sp.
C-153339, P3091-24b, GSC 96354.

Figure 3. *Lonicera* sp.
C-153340, P3091-25b, GSC 96355.

Figure 4. *Pachysandra* sp.
C-163151, P3091-36b, GSC 96356.

Figure 5. *Salix* sp.
C-153342, P3091-27c, GSC 96357.

Figure 6. *Fraxinus* sp.
C-163151, P3091-36c, GSC 96358.

Figure 7. *Fraxinus* sp.
C-163151, P3091-36c, GSC 96359.

Figure 8. *Quercus* sp.
C-163151, P3091-36c, GSC 96360.

Figure 9. *Quercus* sp.
C-153319, P3091-4c, GSC 96361.

Figure 10. *Quercus* sp.
C-153346, P3091-31b, GSC 96362.

Figure 11. *Quercus* sp.
C-163151, P3091-36c, GSC 96363.

Figure 12. *Acer* sp.
C-153329, P3091-14a, GSC 96364.

Figure 13. *Acer* sp.
C-153321, P3091-6b, GSC 96365.

Figure 14. *Acer* sp.
C-153338, P3091-23a, GSC 96366.

Figure 15. *Acer* sp.
C-153331, P3091-16i, GSC 96367.

Figure 16. *Tricolporopollenites kruschii*
C-163172, P3091-57i, GSC 96368.

Figure 17. *Tricolporopollenites kruschii*
C-163158, P3091-43f, GSC 96369.

Figure 18. *Nyssa* sp.
C-163151, P3091-36c, GSC 96370.

Figure 19. *Nyssa* sp.
C-163185, P3091-70c, GSC 96371.

Figure 20. *Quercus* sp.
C-163151, P3091-36c, GSC 96372.

Figure 21. *Fagus* sp.
C-153342, P3091-27j, GSC 96373.

Figure 22. *Fagus* sp.
C-153317, P3091-2a, GSC 96374.

Figure 23. *Tricolporopollenites* sp. ?Acanthaceae
C-163159, P3091-44e, GSC 96375.

Figure 24. *Tricolporopollenites* sp. ?Acanthaceae
C-163160, P3091-45e, GSC 96376.

Figure 25 ?Rosaceae
C-163151, P3091-36a, GSC 96377.

Figure 26. *Castanea* sp.
C-153339, P3091-24c, GSC 96378.

Figure 27. *?Rhus* sp.
C-163151, P3091-36c, GSC 96379.

Figure 28. *Tricolporopollenites* sp.
C-153331, P3091-16j, GSC 96380.

Figure 29. *Tricolporopollenites* sp.
C-153319, P3091-4c, GSC 96381.

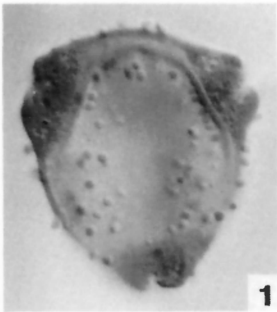
Figure 30. *Tricolpites* sp.
C-163173, P3091-58b, GSC 96382.

Figure 31. *Monocolpopollenites* sp. *?Magnolia*
C-153342, P3091-27i, GSC 96383.

Figure 32. *Liliacidites* sp.
C-153331, P3091-16j, GSC 96384.

Figure 33. *Liliacidites* sp.
C-153331, P3091-16j, GSC 96385.

Figure 34. *Liliacidites* sp.
C-153331, P3091-16h, GSC 96386.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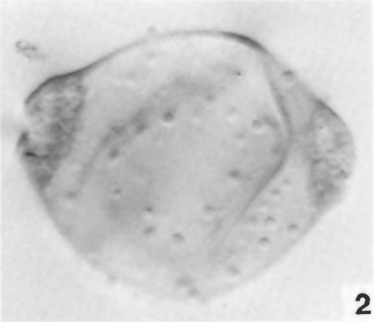
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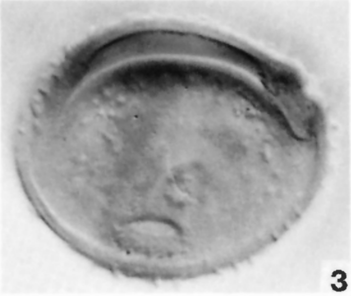
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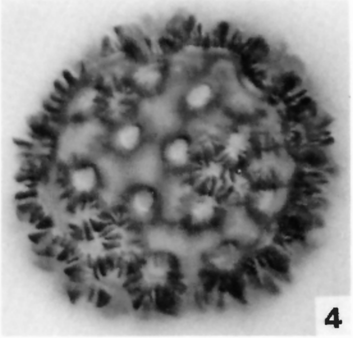
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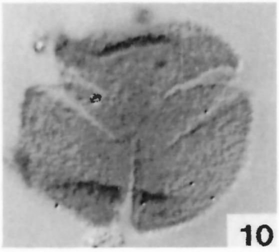
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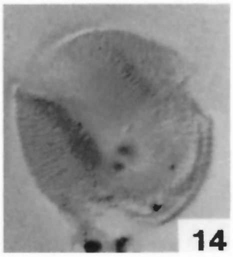
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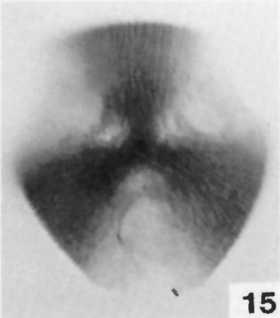
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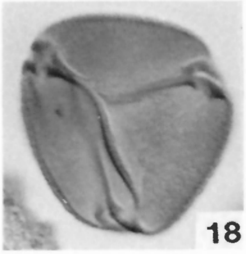
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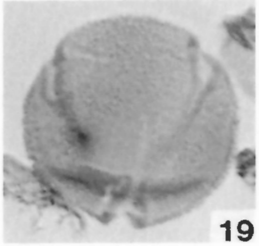
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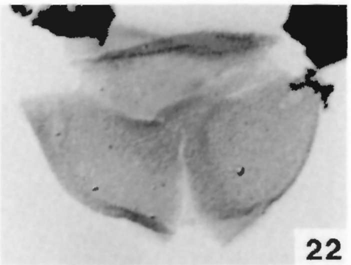
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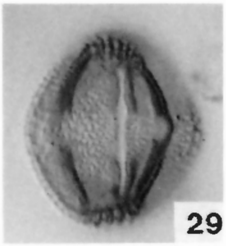
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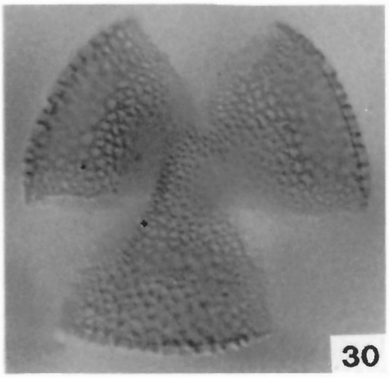
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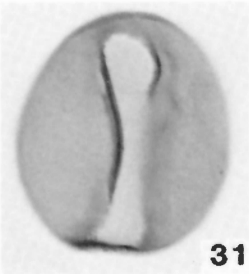
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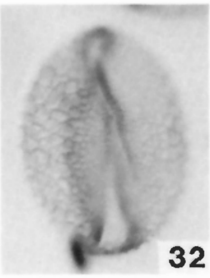
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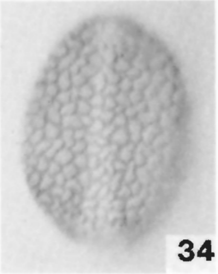
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AN OVERVIEW OF FOSSIL FUNGI IN THE GEODETIC HILLS FOSSIL FOREST, AXEL HEIBERG ISLAND, N.W.T.

Richard G. Day[1]

Day, Richard G., 1991. An overview of the fossil fungi in the Geodetic Hills fossil forest, Axel Heiberg Island, N.W.T. In Tertiary Fossil Forests of the Geodetic Hills, Axel Heiberg Island, Arctic Archipelago, R.L. Christie and N.J. McMillan (eds.); Geological Survey of Canada, Bulletin 403, p. 99-121.

Abstract

This progress report of preliminary laboratory analyses illustrates the wide morphological diversity of the dispersed fungal spores and propagules recovered from leaf mats at several stratigraphic levels. The paleomycological association is of potential use in the interpretation of the paleoecology of the fossil forest.

Résumé

Ce compte rendu des analyses préliminaires de laboratoire illustre la vaste diversité morphologique des spores et propagules fongiques dispersés, qui ont été récupérés dans les tapis de feuilles à plusieurs niveaux stratigraphiques. L'importance potentielle de l'association paléomycologique pour l'interprétation paléoécologique de la forêt fossile est brièvement examinée.

INTRODUCTION

The discovery of the lower Tertiary fossil forest northeast of the Geodetic Hills of Axel Heiberg Island (Fig. 1) in 1985 (Basinger, 1986; McMillan, 1986; Francis and McMillan, 1987; Christie, 1988) presents a unique opportunity for the investigation of fossil fungal assemblages, their relationships, and paleobiological importance. Even though this forest (in reality one unit of several successive ancient forests) is best known for its concentration of in situ, well preserved and non-petrified, erect tree stumps throughout several stratigraphic horizons, significant undisturbed forest-floor leaf-litter mats occur in several layers. Some of these mats are associated with stump-bearing horizons – others are not. The leaf mats vary in botanical composition and consist of shed foliage (leaves and needles), cones, seeds, rootlets and wood fragments, etc. – all the constituents that one might expect in the leaf litter of extant forest communities. The forest floor broadly reflects the forest ecosystem from which the debris was derived.

The paleoflora is taxonomically diverse. Genera represented are *Metasequoia*, *Glyptostrobus*, *Larix*, *Pinus*, *Picea*, *Carya*, *Platanus*, *Abies*, *Alnus*, *Betula*, *Ginkgo*, and *Osmunda*. These archaic leaf mats of the broad-leaved angiosperm and gymnosperm forests are excellently preserved, albeit slightly lignified and compressed. They differ only slightly from well-aged modern leaf litters. The leaf mats in some of the horizons may represent the "A" zone of forest podzols, the typical soil group that develops on cool and moist coniferous forest floors (Francis and McMillan, 1987).

FUNGI

Fungi are obligate or facultative heterotrophs for carbon, an attribute that has resulted in a close evolutionary or co-

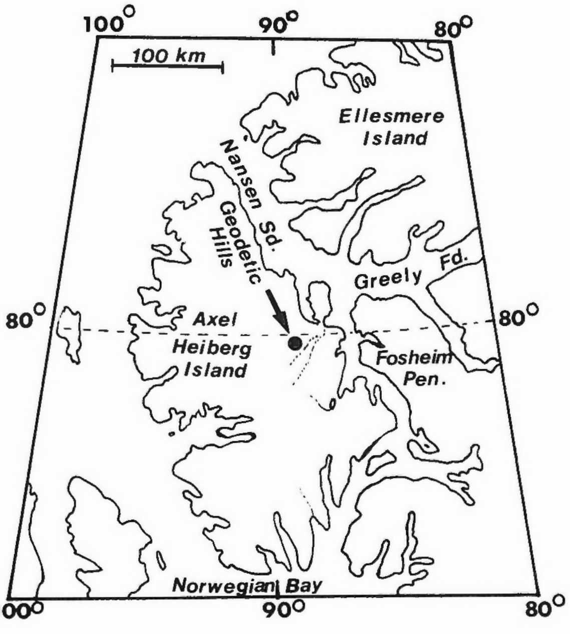

***Figure 1.** General locality of the Geodetic Hills Tertiary fossil forest, Axel Heiberg Island, Canadian Arctic Archipelago.*

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[1] Paleobiology Division, National Museum of Natural Sciences, Ottawa , Ontario K1P 6P4

evolutionary association with plants and animals. Fossil fungi can provide critical knowledge for the interpretation of habits and environmental-ecological parameters of their hosts (Graham, 1962; Elsik, 1976a,b; Pirozynski, 1976; van Geel, 1972, 1986; Pirozynski et al., 1984, 1988). Fossil fungi are mainly derived from two sources: 1) "autochthonous", i.e., fungi that occur in situ, in or directly associated with plant and animal macrofossils, and 2) "dispersed", i.e., *sporae dispersae* in sedimentary and organic matrices. Fungi from an autochthonous source provide more useful phylogenetic-ontogenetic and paleoclimatic data because of their natural association with host tissues and the concomitant knowledge that the host imparts. Dispersed spores, on the other hand, are inherently less informative because of the possibility of postdepositional transport and difficulties in their identification and correlation with extant taxa. Exceptionally, however, some unique and distinctive detached fungal spores could be of significant use as index fossils in biostratigraphy and as indicators of paleoenvironment (Graham, 1962; Kalgutkar, 1985; Kalgutkar and Sweet, 1988).

Despite the recognized pitfalls in their identification and interpretation, dispersed fungal spores in Quaternary lacustrine and fluviatile sediments can be of value in defining a wide range of ecological parameters such as plant succession, temperature regimes and fluctuations, moisture regimes, burning, freshwater/marine inundations, soil formation, grazing pressure, mammalian diet and habitat, etc. (Wolf, 1966a,b, 1967a,b,c,d, 1968, 1969; Wolf and Cavaliere, 1966; van Geel, 1972, 1978; Pals et al., 1980; Davis et al., 1984; Pirozynski et al., 1984, 1988; Jarzen and Elsik, 1986; van Geel et al., 1986).

Even though the primary use of fossil fungi in paleohabitat interpretations has been, especially in more recent studies, for relatively young paleoenvironments, the potential for similar use in much older geological strata exists, particularly in combination with palynological and paleobotanical studies. However, analogies to modern taxa and their respective ecological roles must be developed cautiously.

The mycological study of the fossil leaf mats will be dealt with in two stages:

1. An initial investigation of the dispersed fungal spores of the litter mats, to ascertain whether: a) there is any correlation with palynological microfossil and/or paleobotanical macrofossil remains of a specific mat; b) the fungi or any fungal elements correlate with changes in vegetation and related climatic changes; and c) any of the fungal forms have stratigraphic significance.
2. An investigation of the plant macrofossils (shed foliage, seeds, cones, rootlets, and other structures) from in situ fungal occurrences, particularly for saprophytic or parasitic fungi: leaf or foliicolous ascomycetes, leaf-litter hyphomycetes, wood- and root-inhabiting fungi. The relationship of these macrofossils to modern taxa and the accompanying biological information need to be evaluated.

The purpose of this paper is to illustrate the morphological diversity of the fungal propagules found in a preliminary survey of the fossil forest litter mats.

STRATIGRAPHIC POSITION AND AGE

The collections reported here are from about the middle of a sandstone-siltstone-mudstone lithofacies (the Sandstone Association; see Ricketts, *this volume*) of the Buchanan Lake Formation (Ricketts and McIntyre, 1986). The succession contains stumps buried in growth position. The organic layers, especially of leaf litter, are also widespread in the area surrounding the fossil forest. Leaf-mat beds are well exposed on a higher hill east of the fossil forest site and on a ridge to the south (informally called the "East Hill" and "South Hill", respectively; see Fig. 2).

The fungi illustrated (Pl. 1-6) came from four relatively closely spaced leaf-litter horizons on the south side of the fossil forest hill (Figs. 3, 4). These leaf-litter layers were named beds I, K, M and N by Francis and McMillan (1987).

The Buchanan Lake Formation of the Eureka Sound Group has been assigned a mid-Eocene age by McIntyre (*this volume*) on the basis of palynological data (see also Ricketts

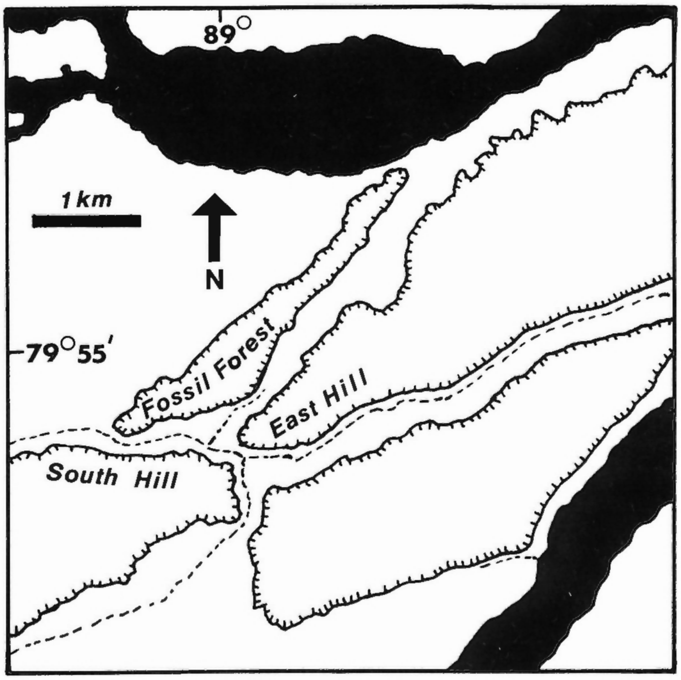

***Figure 2.** Fossil forest site. Heavy shading represents glacial meltwater runoff from the Müller Glacier to the west. Dashed lines represent streams.*

and McIntyre, 1986). The possibility of a late Eocene age for the macroflora of this formation is discussed by Basinger (*this volume*). Bustin (1982) had earlier considered the fossil forest beds to be of Miocene or younger age.

FOSSIL SAMPLES

The leaf-litter layers from which the fungal remains (Pl. 1-6) were recovered represent a relatively narrow portion of the measured section of Francis and McMillan (1987). Samples for paleomycological study were collected in 1987 from the most promising of the forest-floor litters. The leaf mats vary in thickness from approximately 10 cm to 50 cm.

Some of the mats can be differentiated on the basis of their macrofossil assemblages, reflecting distinct vegetational changes within the same forest-floor unit.

The litter mats studied so far (expressed in the stratigraphic nomenclature of Francis and McMillan, with minor modification) can be characterized chronologically from older to younger units as follows:

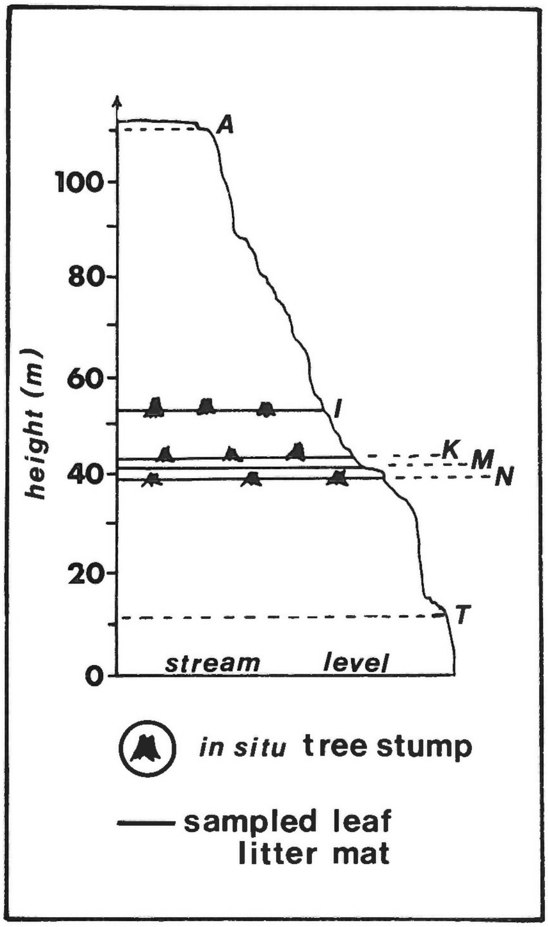

Figure 3. Partial section of the Geodetic Hills fossil forest showing approximate levels of the fungal-bearing leaf mats. (Adapted from Francis and McMillan, 1987.)

Level N(b) (Paleomycology microslide prefix AHI-2.)

This is the lower part of this level. N(b) is a thin, well-preserved litter mat almost monospecific in nature with *Metasequoia* dominating, in contrast to the upper part of level N where *Metasequoia* is mixed with a variety of broad-leaved angiosperms.

Level M (Paleomycology microslide prefix AHI-3.)

This is a coarse litter horizon, mainly of *Metasequoia* foliage, about 15 cm thick and containing abundant rootlet and wood fragments.

Level K(b) (Paleomycology microslide prefix AHI-4.)

This is a well-preserved mixed leaf litter, 10 to 20 cm thick, at the bottom of level K, consisting mainly of *Alnus* and *Osmunda*, whereas the upper portion of this level is a mixture of *Alnus*, *Carya*, *Gingko*, *Betula* and *Osmunda*.

Level I (Paleomycology microslide prefix AHI-1.)

This is a relatively poorly preserved unit of carbonized "woody peat", principally of *Metasequoia* (needles and cones), *Larix*, and *Alnus*, and is about 10 to 20 cm thick.

Most fungal remains were recovered from the organic residue of horizons M and I, which is perplexing in view of the coarseness and poorer condition of these samples. The lower diversity of fungal remains at N(b) might be anticipated because of the almost monospecific nature of this level. The diversity of fungal taxa encountered at level K(b) was similar to that of N(b), despite the former's more diverse vegetation.

METHODS

All samples were processed as uniformly as possible, at least until after acid-oxidation, in order to minimize any processing bias that would affect the recovery of fungal propagules. Acid-oxidation of a sample followed by base treatment can result in the loss of fungal spores if sample residues are over-oxidized. Standard palynological techniques for processing peat, with minor modifications, were used in the following sequence:

1. A pulverized 10 g sample of each leaf-litter mat was heated for 5 minutes at 90°C in 5% KOH, followed by repeated water washing. (The water washes are a standard step after and between any reagent treatment.)
2. The sample was subjected to 5 minutes of ultrasonic vibration in a Bransonic 12 ultrasonic cleaner, in

order to facilitate the release of spores from adhering cuticles or mineral matter. Although ultrasonic treatment may result in damage or disruption to fungal spores and other palynomorphs, no appreciable deterioration was observed.

3. The sample was then subjected to treatment in cold 48% HF for 24 hours to remove siliceous components followed by three washes in hot 25% HCl to remove any chemical byproducts of the HF reaction.

4. A one-time oxidation of the organic residue involved 2 hours immersion in concentrated HNO_3, followed by separate alkali treatments to remove humic compounds present after oxidation, initially in 5% K_2CO_3 and, if necessary, followed by treatment with 5% NH_4OH, if the amount of extraneous fine grained humic acids remained high after several treatments in the 5% K_2CO_3 reagent. The number of base treatments varied for each sample, depending upon the concentration of humic acids remaining after oxidation.

5. Heavy-liquid separation ($ZnBr_2$ of specific gravity 1.95) using the float/sink technique was employed for samples that required removal of heavy minerals remaining after HF treatment.

6. Owing to the volume of humic acids still present after the alkali treatments, each sample was washed three to four times with Darvan no. 4 to aid in the dispersion and removal of impurities.

7. The organic residue was stained in a 1% solution of SAF-O, mainly to differentiate pollen and other spores from the fungal palynomorphs; the melanized fungal spores do not readily absorb the stain.

8. The organic residue was separated into three components using Buckbee-Mears Co. BMC stainless steel micromesh sieves (nominal mesh sizes of 20 and 90 μm). A standard volume of each fraction was mounted on microscope slides with Cellosize WP-4 and Harleco Krystalon mounting media. The residue was separated into three fractions because of the abundance of large organic particles and the quantity of pollen and fungal spores. Separation facilitates microscopic examination. Excess residues were stored in vials in water with a few drops of phenol added to discourage growth of fungal contaminants.

Prepared slides and residues are stored at the Paleomycology Section of the Paleobiology Division of the National Museum of Natural Sciences (Ottawa) under the general prefix "AHI."

Figure 4*. View east across the southern exposures of the fossil forest toward the prominent "East Hill" ridge. The arrow indicates the leaf mat horizons from which samples were collected for this report.*

FOSSIL FUNGAL SPORES

The dispersed fungal spores encountered in the fossil forest leaf mats are mainly members of the largest group of fungi, the Ascomycotina, and the group Fungi Imperfecti (Deuteromycotina), in which members are generally anamorphic and reproduce mainly by means of conidia. Although the majority of the spores in the Fungi Imperfecti are presumed to be asexual stages of some ascomycetes (sexual or teleomorph state), it could be that some fungi propagate only asexually, having either lost their ability to reproduce sexually, or their sexual stages have yet to be identified. For terms and details of classification of the fungi, readers are referred to Hawksworth et al. (1983).

CLASSIFICATION AND NOMENCLATURE OF FOSSIL FUNGI

Many of the dispersed spores from the leaf mats resemble Tertiary fungal propagules illustrated elsewhere (Dilcher, 1965; Elsik, 1968; Lange and Smith, 1971; Sheffy and Dilcher, 1971; Elsik and Dilcher, 1974; Elsik and Jansonius, 1974; Jansonius, 1976; Rouse, 1977; Lange, 1978a,b; Smith, 1978; Ioannides and McIntyre, 1980; Ediger, 1981; Sepulveda and Norris, 1982; Choi, 1983; Norris, 1986; Dietrich et al., 1989). A number of these documented fungi have been incorporated into form-morphographic categories, originally devised as cataloguing schemes for fossil fungi by palynologists using palynological terminology, as compared to the morphographic "Saccardoan" system used by mycologists, which facilitates correlation between fossil and extant fungal forms. The utilitarian and philosophical implications of such information-retrieval systems have been discussed by Elsik (1976a,b) and Elsik et al. (1983), who adopted the former, and Lange (1978a) and Pirozynski and Weresub (1979), who advocated the latter. Smith (1981) discussed both approaches. He identified further problems complicating the circumscription of fossil fungal biotas, particularly at the generic level, due to an inherent variability in fungal spore morphology (size, shape, septation) correlated with external factors such as host/substrate chemistry, host age and temperature regimes.

In this report, dispersed fungal spores from the fossil forest are classified according to the informal Saccardoan system. As Pirozynski and Weresub (1979) have pointed out, "undetermined fossil spores placed alongside spores of similar morphology but from living fungi will have the best possible chance of being identified or judged as belonging to an extinct taxon. Or if not, if they are so poorly preserved or so lacking in distinctive characters that the fossil entity cannot be given taxonomic standing need it be classified any further?"

The Saccardoan system has been particularly useful for classifying conidia. However, for this investigation, both conidia and ascospores are included. (A primary feature distinguishing conidial structures from ascospores is the presence of an attachment scar on the former.) The Saccardoan system also has its shortcomings, as discussed by Kendrick and Nag Raj (1979), who examined the utility of the system, and modified it with requisite but arbitrary limitations. The Saccardoan system as employed here has been simplified from that in Hawksworth et al. (1983) and as used by Lange (1978a) for the classification of Tertiary fungal spores from Australia. This system, as used to categorize the Geodetic Hills *Sporae dispersae*, is shown in Table 1.

Figure 5 depicts an idealized graphic representation of the different morphographic groups of the Saccardoan system. Apart from the general shape of a fungal spore, the Saccardoan system incorporates the pigmentation of the cell wall, i.e., it segregates nonpigmented (hyaline) and melanized spores. As all the fungal spores encountered in our samples so far appear to have been melanized, this aspect of the system is arbitrarily ignored. Nonsporic components such as bulbils, sclerotia, mycelial mats, and hyphae, although present in the samples, have generally not been taken into consideration. Examples of the latter are given in Plates 1 to 6 and the accompanying plate descriptions. No apparent scolecosporous spores have, as yet, been documented in my examination.

FOSSIL FOREST FUNGAL SPORES

Overall, the dispersed spores as shown in the plates have a terrestrial "flavour" (K.A. Pirozynski, pers. comm.). Many of

TABLE 1

Saccardo's spore groups[1,2]

Spores		Category
A	1-celled (nonseptate)	Amerosporae
B	2-celled (one-septate)	Didymosporae
C	2- to many-celled [multiseptate in one plane (horizontal)] having a length/width ratio not exceeding 15:1	Phragmosporae
D	Muriform (multiseptate with both vertical and horizontal septa)	Dictyosporae
E	Filiform having a length/width ratio exceeding 15:1	Scolecosporae
F	Helical or spirally coiled	Helicosporae
G	Stellate or star-like in form	Staurosporae

[1] Adapted from Hawksworth et al. (1983), Kendrick and Nag Raj (1979) and Pirozynski and Weresub (1979).

[2] See Figure 5 for idealized graphic interpretation of the Saccardoan spore morphographic groups.

these forms, particularly conidia, compare with those of living fungi as depicted in Ellis (1971, 1976). For example, a *Sporidesmium/Clasterosporium* affinity is suggested for conidia in Pl. 4, figures 120 to 131. Members of both these genera are cosmopolitan and occur on a wide variety of dead wood and leaves. In cases such as these, the generic rank provides little data on paleoenvironment. Identification to species level would be more advantageous for paleoecological data refinement.

Some of the amerosporous ascospores (Pl. 1, figs. 7–33) may be referable to modern Sordariales, a group of fungal saprobes found in dung, rotting vegetation and wood, and soil—some are even myco-parasites. Other propagules may be part of a *Brachysporiella-Endophragmia-Bactrodesmium* complex (Pl. 3, figs. 87–91, 96–104, 108, 110–112, 117), fungi that are characteristically saprobes of leaves and bark of broad-leaved angiosperms and herbaceous plants, typically found in, although not restricted to temperate climates. However, any correlations of these fossil forms to the living biota are premature without full reference to the literature and comparative herbarium material.

Other fossil remains illustrated in the plates can be referred to taxa known from the fossil record. Figures 174 and 175 (Pl. 6) are but two examples of an abundant *Ctenosporites* component that occurs in the Geodetic Hills samples. *Ctenosporites*, apparently confined stratigraphically to Tertiary sediments (Eocene-Oligocene), is known only from deposits in the Canadian northwest Pacific and Arctic regions, South Australia and southern England (Elsik and Jansonius, 1974; Rouse, 1977; Lange, 1978a,b; Smith, 1978; Norris, 1986; Dietrich et al., 1989). According to Smith (1978), spores of *Ctenosporites eskerensis* are thought to be related to the conidia of the living saprophytic hyphomycete *Dictyosporium toruloides*. In this study, *Ctenosporites* remains have been recovered from levels K(b) and M.

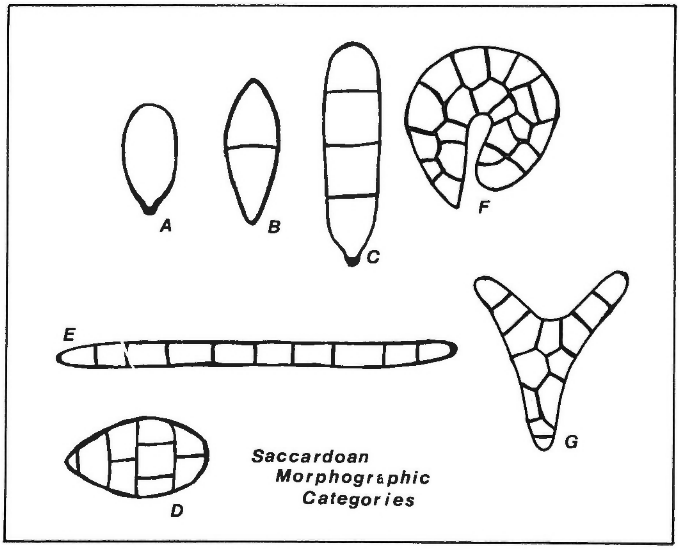

***Figure 5**. Graphic representation of the Saccardoan morphographic categories: A, Amerosporae; B, Didymosporae; C, Phragmosporae; D, Dictyosporae; E, Scolecosporae; F, Helicosporae; and G, Staurosporae. (Modified from Kendrick and Nag Raj, 1979, and Hawksworth et al., 1983.)*

Other fungal propagules recovered from the fossil forest leaf mats can tentatively be compared to fossil genera that have been documented and described from Paleogene sediments of the Canadian Arctic (Jansonius, 1976; Rouse, 1977; Ioannides and McIntyre, 1980; Sepulveda and Norris, 1982; Choi, 1983; Norris, 1986; Dietrich et al., 1989); for example: *Fusiformisporites* (Pl. 2, fig. 42), *Punctodiporites* (Pl. 6, fig. 166), *Dicellaesporites* (Pl. 2, figs. 45, 46, 49) and *Striadiporites* (Pl. 1, figs. 2, 3), a taxa confined in Arctic Canada to the Eocene (Rouse, 1977; Sepulveda and Norris, 1982).

The presence of numerous amerosporous ascospores (Pl. 1, figs. 1–3 (cf. *Striadiporites*)) is notable in the Axel Heiberg samples. These amerospores have so far been stratigraphically restricted to the M horizon.

Undoubtedly, with further investigation, it will be possible to refer more of the leaf-mat fungal propagules to either living taxa or to fossil forms already documented in the palynological literature.

SUMMARY

As is apparent from Plates 1 to 6, there is a very diverse and abundant complement of fungal propagules in the fossil forest leaf mats. The fungi shown represent either those taxa with distinctive morphological features and/or the more commonly encountered forms.

In the initial phases of this investigation, it was expected that fossil fungi would be found associated with the leaf mats because fungal spores are usually fairly abundant in sediments or matrices with a large component of organic matter, such as wood and cuticular fragments, presumably as a reflection of saprophytic fungi at work. The fungi have a long geological record, probably dating to the Precambrian. During the Cenozoic, they evolved rapidly, most likely linked to the evolution and diversification of angiosperms. It was both surprising and rewarding to find such a wide variety of fungal forms in these Eocene leaf litters, and, in relatively high proportions compared with other palynomorphs, such as pollen.

The fungal propagules, like the megafossils of the forest, are well preserved and show little evidence of geothermal alteration, or other forms of degradation. In general, chitinouswalled fungal spores are reasonably resistant to damage; however, as Elsik is quoted in Traverse (1988), "fungal spores may be either more or less resistant to both carbonization and oxidation, depending on as yet poorly understood factors." In this study, it is not known why there should be a differential recovery in fungi, with a more numerous and richer variety of fungi found in the more poorly preserved and less botanically diverse horizons.

Some of the fossil fungal taxa have been compared to living forms. These extant fungi can provide either general or specific data as indicators to paleohabitat. Other fossil propagules have been referred to previously known fossil fungi. Some of these spores show regional distribution whereas others are widespread. Some show a wide geological age range, and several fungal propagules of unique morphology and a restricted stratigraphic range have been useful as index fossils in biostratigraphic correlations. The possibility exists that some of the other fossil taxa of the fossil forest may have a similar usefulness. Some fungi are known to be "host specific." Could any of the fungi of the fossil forest provide clues as to the plant or animal fossils that may have existed there but which have not left any fossilized evidence of their existence?

This progress report generates more questions than answers concerning the role of fungi in paleohabitat reconstructions. The uncertainties inherent in the ecology, life history and classification of the fungi, especially the Ascomycotina (Hawksworth, 1985), magnified in their fossil counterparts, compounds the search for answers. Even acknowledging that the fossil fungi play a subordinate role to pollen (Pirozynski et al., 1988) and plant macrofossils in stratigraphic and paleoenvironmental interpretations, any additional data will add refinement to paleobiological reconstruction.

The Geodetic Hills fossil forest offers an excellent opportunity to investigate any possible qualitative-quantitative relationships between fungal microfossils and other palynomorphs and megafossils co-occurring at the same leaf-litter horizon and also through a sequence of such horizons.

ACKNOWLEDGMENTS

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PLATES

All photographs were taken using a Nikon Optiphot microscope. All photographs x1000, unless otherwise indicated.

PLATE 1

Figures 1-33.	Amerosporae : ascospores.
Figures 34-37.	Amerosporae : conidia.
Figures 38,39.	Dictyosporae : ascospores.
Figures 40,41.	Dictyosporae : conidia.

PLATE 1


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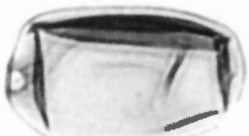
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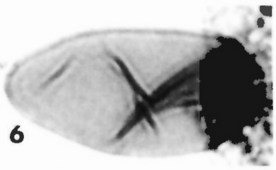
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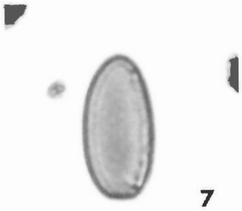
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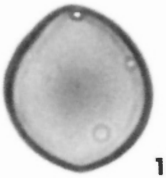
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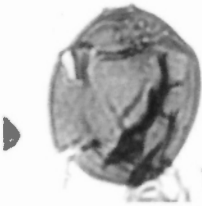
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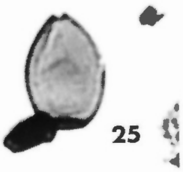
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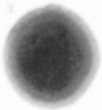
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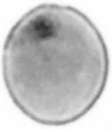
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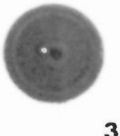
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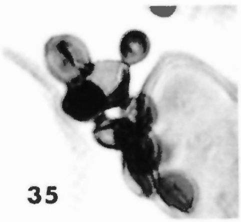
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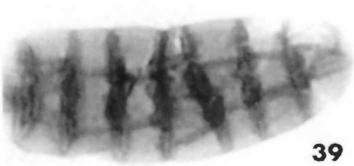
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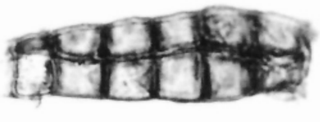
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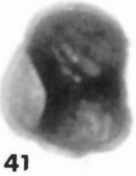
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PLATE 2

Figures 42-72.	Didymosporae : ascospores.
Figures 73-78.	Didymosporae : conidia.
Figures 79-81.	Part-spores, affinity unknown.

PLATE 2


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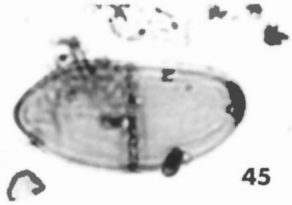
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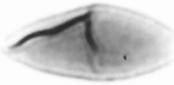
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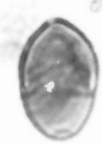
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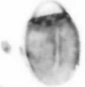
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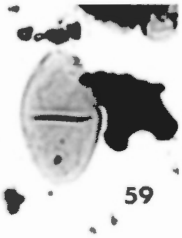
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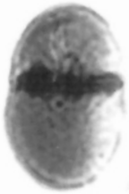
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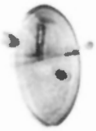
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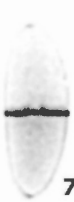
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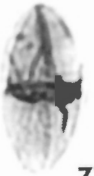
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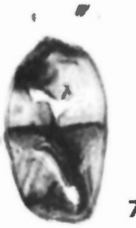
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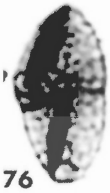
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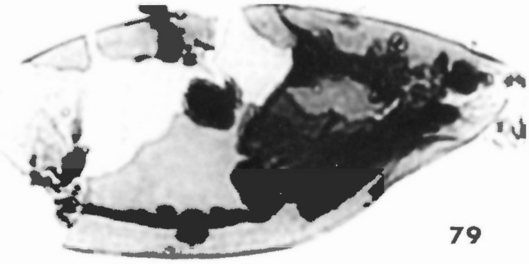
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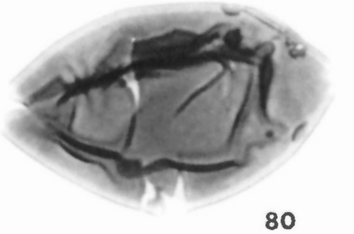
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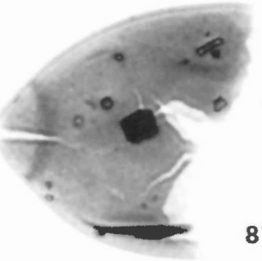
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PLATE 3

Figures 82-117.	Phragmosporae : conidia.
Figure 118.	Phragmosporae : conidium.
Figure 119.	Phragmosporae : conidium (x500).

PLATE 3


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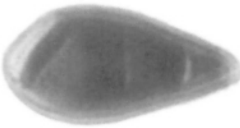
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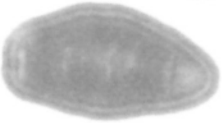
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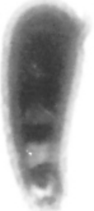
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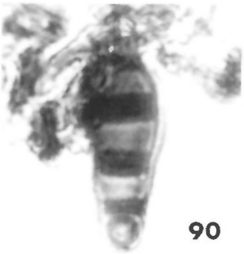
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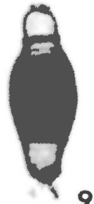
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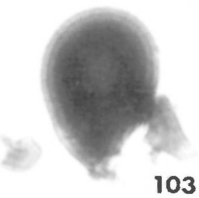
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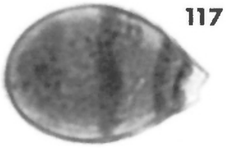
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PLATE 4

Figures 120-125. Phragmosporae : conidia (x500).

Figures 126-136. Phragmosporae : conidia.


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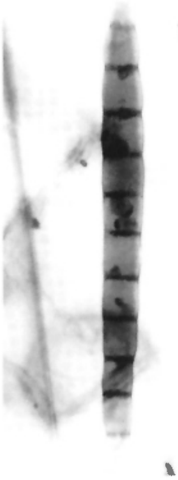
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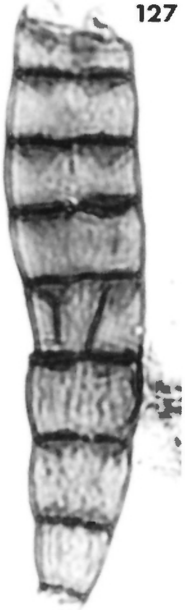
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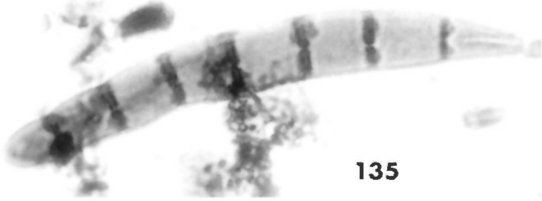
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PLATE 5

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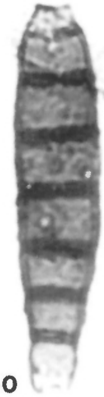
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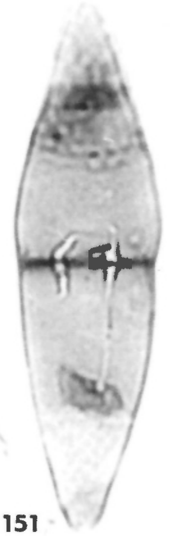
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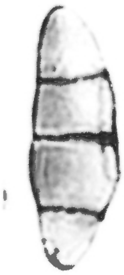
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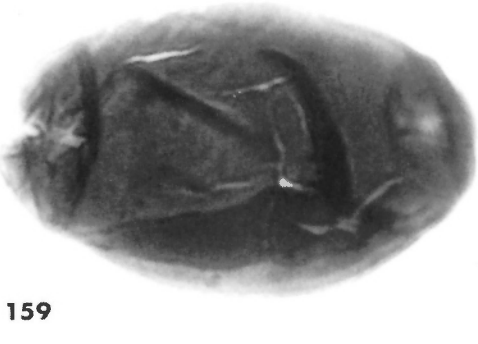
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PLATE 6

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PLATE 6


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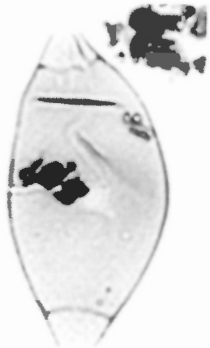
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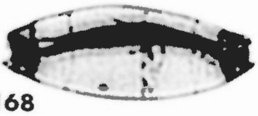
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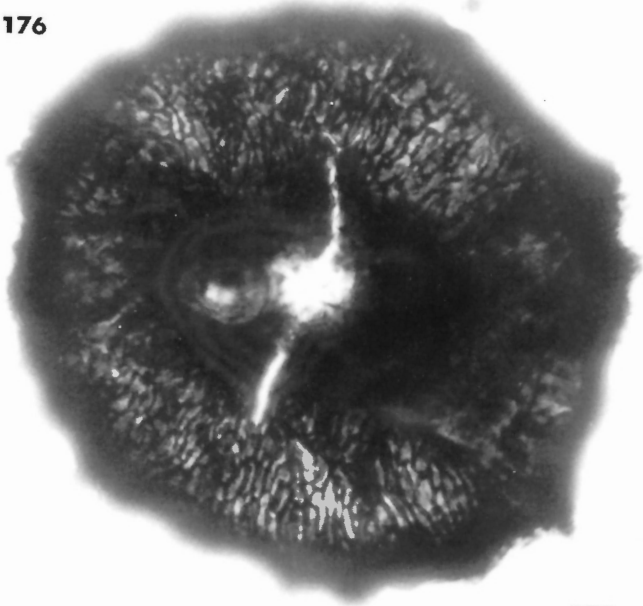
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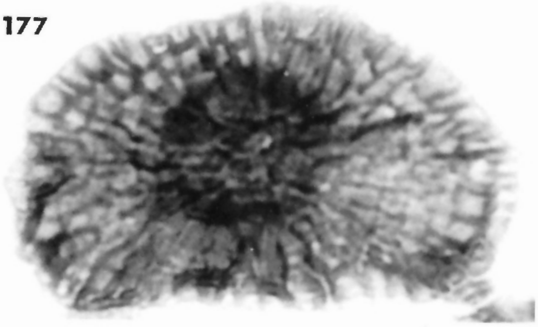
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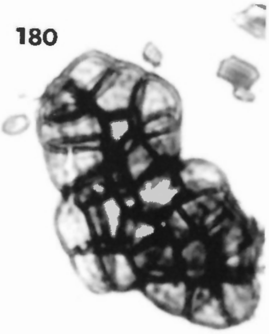
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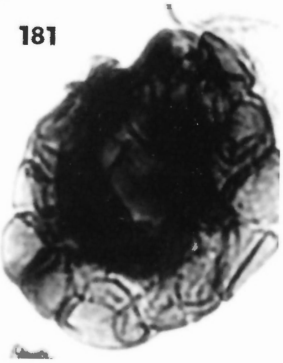
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CHARACTERIZATION OF CANADIAN ARCTIC FOSSIL WOODS

**J.R. Obst[1], N.J. McMillan[2], R.A. Blanchette[3], D.J. Christensen,[1]
O. Faix[4], J.S. Han[1], T.A. Kuster[1], L.L. Landucci[1], R.H. Newman[5],
R.C. Pettersen[1], V.H. Schwandt[1], and M.F. Wesolowski[1]**

Obst, J.R. et al., 1991. Characterization of Canadian Arctic fossil woods. In Tertiary Fossil Forests of the Geodetic Hills, Axel Heiberg Island, Arctic Archipelago, R.L. Christie and N.J. McMillan (eds.); Geological Survey of Canada, Bulletin 403, p. 123-146.

Abstract

Fossil woods from Axel Heiberg, Ellesmere, and Cornwallis islands in the Canadian Arctic were analyzed. Most specimens have been dated as Eocene, but one may be Paleocene, and as old as 65 million years. All the woods are gymnosperms. Based on recognizable characteristics, some of the Eocene specimens have been characterized as possibly spruce or larch, and one specimen has all the characteristics of Douglas fir. A Miocene wood has been identified as an especially well preserved white pine. Chemical and instrumental analyses showed that the Eocene and Paleocene specimens have undergone extensive carbohydrate degradation with nearly complete removal of hemicelluloses. The residual carbohydrate is mainly crystalline cellulose. The fossil woods are rich in lignin; some specimens contain over 80 per cent lignin. The fossil lignins are methoxyl-group deficient, sidechain degraded, and more condensed compared to recent softwood lignin. No evidence of bacterial or fungal decay was observed; hydrolysis was probably the major route of degradation.

Résumé

On a analysé des bois fossiles provenant de l'île Axel Heiberg dans l'Archipel arctique canadien. La plupart des spécimens ont été datés de l'Éocène, mais l'un d'eux pourrait remonter au Paléocène, et avoir un âge de 65 millions d'années. Tous les types de bois proviennent de gymnospermes. D'après leurs caractéristiques identifiables, quelques-uns des spécimens datant de l'Éocène ont été identifiés comme appartenant peut-être à des épinettes ou à des mélèzes, et l'un des spécimens a toutes les caractéristiques du sapin de Douglas. Un bois d'âge miocène a été identifié comme un échantillon particulièrement bien conservé de pin blanc. Les analyses chimiques et physiques ont montré que les spécimens datant de l'Éocène et du Paléocène ont subi une importante dégradation des hydrates de carbone, avec élimination presque complète des hémicelluloses. L'hydrate de carbone résiduel est principalement de la cellulose cristalline. Les bois fossiles sont riches en lignine; certains spécimens contiennent plus de 80% de cette substance. Les lignines fossiles ont un groupe méthoxyle déficitaire, des chaînes latérales dégradées, et sont plus condensées que les lignines des résineux récents. On n'a observé aucun indice d'une décomposition bactérienne ou fongique; l'hydrolyse a probablement été la principale voie de dégradation.

INTRODUCTION

Arid, barren, and cold, Axel Heiberg Island supports only specialized arctic plants. Located over 1000 km north of the Arctic Circle, its weather is typical of polar climates. Yet this area was the site of lush forests 45 to 60 million years ago (Francis and McMillan, 1987). These conifer forests, which had trees of up to 2.5 m in diameter and possibly over 50 m in height, are thought to have been inundated by floods and ultimately buried under 180 to 300 m of sand and silt. Eventually uncovered by erosion, the Axel Heiberg fossil forests are a source of fascination and a subject for scientific inquiry.

Fossil forests, clusters of stumps with or without attached trunks, have been discovered at a number of sites worldwide. Certain Canadian Arctic sites are notable for their excellent preservation. At the Geodetic Hills site on Axel Heiberg Island, more than 20 forests have been found in stacked layers.

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[1] USDA Forest Service, Forest Products Laboratory*, Madison,WI 53705-2398, U.S.A.

[2] Institute of Sedimentary and Petroleum Geology, Geological Survey of Canada, Calgary, Alberta T2L 2A7

[3] Department of Plant Pathology, University of Minnesota, St. Paul, MN 55108, U.S.A.

[4] Institut für Holzchemische Technologie des Holzes, Hamburg, West Germany

[5] Chemical Division, DSIR, Petone, New Zealand

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Arctic fossil woods were discovered over 100 years ago (Brainard, 1886). A unique feature of the Canadian Arctic fossil forests is that many of the trees have remained woody and are not petrified (permineralized) (McMillan, 1963). Although chemical analyses of ancient woods, which range in age from hundreds and thousands of years (Borgin et al., 1975) to millions of years (Wayman et al., 1972) have been described, no detailed analysis of these Canadian Arctic fossil woods has been reported until now.

This report describes a single wood specimen of Paleocene age from Hot Weather Creek on Ellesmere Island, a suite of 10 specimens of Eocene age from the Geodetic Hills on Axel Heiberg Island, and a single specimen, possibly of Miocene age, from Cornwallis Island.

DESCRIPTION OF SPECIMENS

The source and age of specimens are listed in Table 1. The Miocene wood, here called Resolute, was collected near Resolute Airport on Cornwallis Island. This specimen is the best physically preserved specimen. It is light brown and looks rather like ordinary, weathered wood.

The Eocene specimens were all collected from the fossil forests at Geodetic Hills (GDH), Axel Heiberg Island (lat. 79°55'N, long. 88°58'W). The corresponding Geological Survey of Canada Type Numbers are given in Appendix Table 1. All these specimens are quite dark, compressed, hard, and brittle. Although palynological and paleobotanical evidence supports an Eocene date, at least one botanist suggests that the GDH beds may be Oligocene (L.V. Hills, pers. comm., 1989).

The Paleocene specimen was collected from Hot Weather Creek (HWC) on Ellesmere Island (lat. 79°58'N, long. 84°48'W). This specimen is similar in appearance to the GDH specimens, but somewhat darker and more physically degraded.

TABLE 1

Source and age of specimens

Specimen	Source	Age (years before present)
Paleocene (HWC 37)	Hot Weather Creek Ellesmere Island	up to 65 000 000
Eocene[a]	Geodetic Hills, Axel Heiberg Island	45 000 000 to 55 000 000
Miocene[b]	Cornwallis Island	up to 24 000 000
Spruce	Wisconsin	500 000
Spruce	Michigan	9900
Loblolly pine	North Carolina	(present)

[a] GDH specimens plus NMNS S1.

[b] Resolute.

SPECIES IDENTIFICATION

Based on identification of leaves and cones, the gymnosperms (conifers) of the Axel Heiberg fossil forests have been identified as dawn redwood (*Metasequoia glyptostroboides*) (Francis and McMillan, 1987). However, because this identification was not based on the wood, we examined the Miocene (Resolute) and Paleocene (HWC 37) specimens and seven of the ten Axel Heiberg specimens microscopically. Specimens GDH 40, 44, and 74 were not examined microscopically because of their poor condition. Both the gross condition (degree of abnormal density and distortion) and microscopic condition (degree of cell degradation and presence of extraneous substances) vary considerably among these specimens.

Microscopic characterization

Although a generic identification of the Resolute specimen was made easy by its excellent preservation, we had difficulty in comprehending how this specimen fared so well after being buried for millions of years. By contrast, the HWC 37 specimen is so compressed and distorted that only a broad identification was possible. The Axel Heiberg specimens all exhibit appreciable amounts of cell deterioration. Specimens subjected to such extreme circumstances presumably lose many of the anatomical characteristics commonly present in normal wood. With the exception of the Resolute specimen, this is the case with the fossil wood specimens. Hence, the level of precision in identification depends on the degree to which the original cell structure has remained intact and visible.

All specimens examined are conifers. Even HWC 37, the oldest and most degraded specimen, displays bordered pits (Fig. 1), a feature commonly used to distinguish conifers from hardwoods (angiosperms). Unfortunately, the presence of bordered pits is the only anatomical characteristic clearly evident in HWC 37. Accordingly, we are restricted to broadly classifying this specimen as a conifer.

Conversely, the Miocene specimen, Resolute, exhibits in fine detail all the characteristics that distinguish white pine from other woods (Fig. 2). The most significant features are resin canals with thin-walled epithelium, nondentate ray tracheids, and one (primarily) or two large, window-like (fenestriform) pits for each cross-field of ray parenchyma cell. All other white pine anatomical features are also intact and clearly visible. Present-day species of the white pine group native to North America are *Pinus strobus*, *P. monticola*, *P. flexilis*, and *P. albicaulis*. Anatomical features of the white pine group are so similar that if only the wood is available for examination, species identification is not possible.

Classification of Axel Heiberg specimens

The remaining seven specimens were divided into groups. All of these specimens contain compression wood, which complicates the identification procedure. Compression wood can moderately affect normal cell size, shape, and arrangement, or severely alter normal structures. For the sake of simplicity, identifications were arbitrarily restricted to North American species. If evidence were presented in the future to indicate the likelihood of tree species from other continents, the proposed identifications could be readily reexamined to include additional species.

For descriptive purposes, we divide the specimens into four distinct categories. The first contains specimens GDH 28 and GDH 59, which are identified as either spruce (*Picea*) or larch (*Larix*). The diagnostic features displayed by this group are resin canals with thick-walled epithelium, ray tracheids, and small cross-field pits. However, even under ideal conditions, the characteristics that separate spruce from larch are difficult to see and to evaluate. In these two specimens, these features are either missing or distorted, and therefore we cannot obtain more specific identification.

Specimens GDH 54, GDH 64, and GDH 80 are placed in a second group. The most diagnostic feature of this category is the absence of resin canals. Unfortunately, the degree of degradation of these specimens is such that only outlines of structures remain and the details of the individual cells are missing. For example, ray tracheids are present in some and absent in others. These specimens could be any conifer except pine, spruce, larch, or Douglas fir, but their condition severely limits any more specific identification. Thus, possible North American genera include *Tsuga* (hemlock), *Abies* (fir), *Juniperus* (juniper), *Libocedrus* (incense-cedar), *Cupressus* (cypress), *Taxus* (yew), *Torreya* (torreya), *Thuja* (cedar), *Chamaecyparis* (cedar), *Taxodium* (bald cypress), and *Sequoia* (redwood).

Group three contains the single specimen NMNS S1. This specimen is also severely distorted and degraded. We were fortunate, however, to detect two distinguishing characteristics: ray parenchyma cells with nodular end walls and vertical parenchyma walls with nodular end walls. North American genera that possess these anatomical features are *Tsuga* (hemlock), *Abies* (fir), *Juniperus* (juniper), *Chamaecyparis* (cedar), *Libocedrus* (incense-cedar), *Cupressus* (cypress), and *Pseudotsuga* (Douglas fir).

Specimen GDH 61 forms the fourth category. This specimen, like the specimens in the first group, has resin canals with thick-walled epithelium, ray tracheids, and small cross-field pits. In addition, spiral thickenings are present, which is indicative of Douglas fir (*Pseudotsuga*) (Fig. 3).

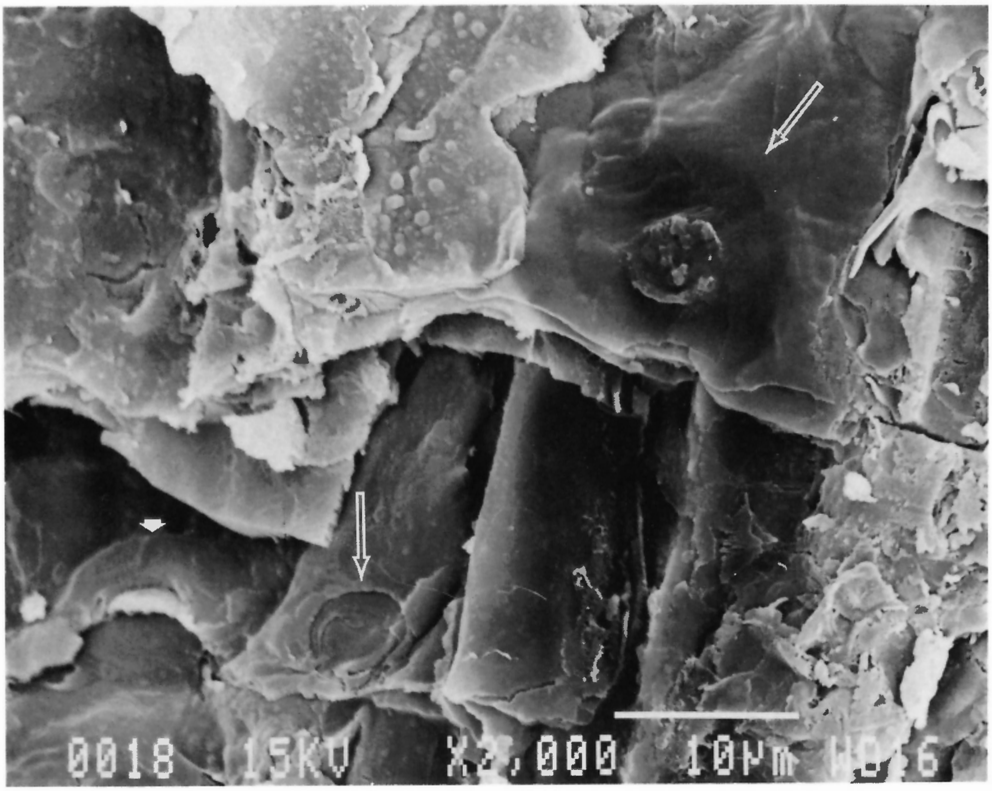

Figure 1*. Scanning electron micrograph of the Paleocene fossil wood HWC 37 from Ellesmere Island. Although impressions of bordered pits (arrows) and a partial bordered pit (arrowhead) can be clearly seen, this specimen is very degraded and could only be identified as a conifer.*

For perspective, the major anatomical features of dawn redwood (*Metasequoia*) are: 1) smooth end walls of parenchyma cells in both the rays and the vertical parenchyma; 2) taxodioid cross-field pits; and 3) absence of resin canals, ray tracheids, and spiral thickenings. [Redwood (*Sequoia*) also has these features.] Thus, of the specimens examined, HWC 37 and GDH 54, 64, and 80 could possibly be dawn redwood. Of course, these specimens could also belong to any of a number of other genera.

Particularly intriguing are the Douglas-fir-like characteristics of GDH 61. Douglas fir is generally restricted to moist Pacific Coast forests and does not commonly occur in northern spruce-pine-larch forests. However, we must recall that our identifications are from the perspective of North American species present today. For example, about 40 species of sequoias existed in North America 60 million years ago (Brockman, 1968). Now, depending upon classification, only one or two species survive (Brockman, 1968). Thus, specimen GDH 61 is best characterized by defining its wood as Douglas-fir-like and recognizing that the specimen could be an extinct species. Consequently, the woods identified as white pine, spruce, and larch could also represent extinct species.

ULTRASTRUCTURAL STUDIES

To describe the ultrastructural aspects of degradation, the fossil woods were examined by light and electron microscopy (see Appendix). Further, to help understand the degradation of buried arctic fossil woods, we also examined buried spruce specimens from Michigan and Wisconsin (Table 1). Thick sections of the 9900-year-old spruce showed some apparently sound cells (Fig. 4A, arrowheads) and degraded adjacent cells. Areas of sound cells showed pit apertures with intact but aspirated pit membranes. Degraded cells appear to have no S_3 or S_2 layers (Fig. 4A). The compound middle lamella has remained intact but has lost its normal rigid structure. Altered cells are collapsed, causing the cell wall to bend or become convoluted. Electron-dense deposits (Fig. 4B, asterisks) were found in varying amounts on parts of the altered cell walls. (These electron-dense deposits are likely to be degraded wood

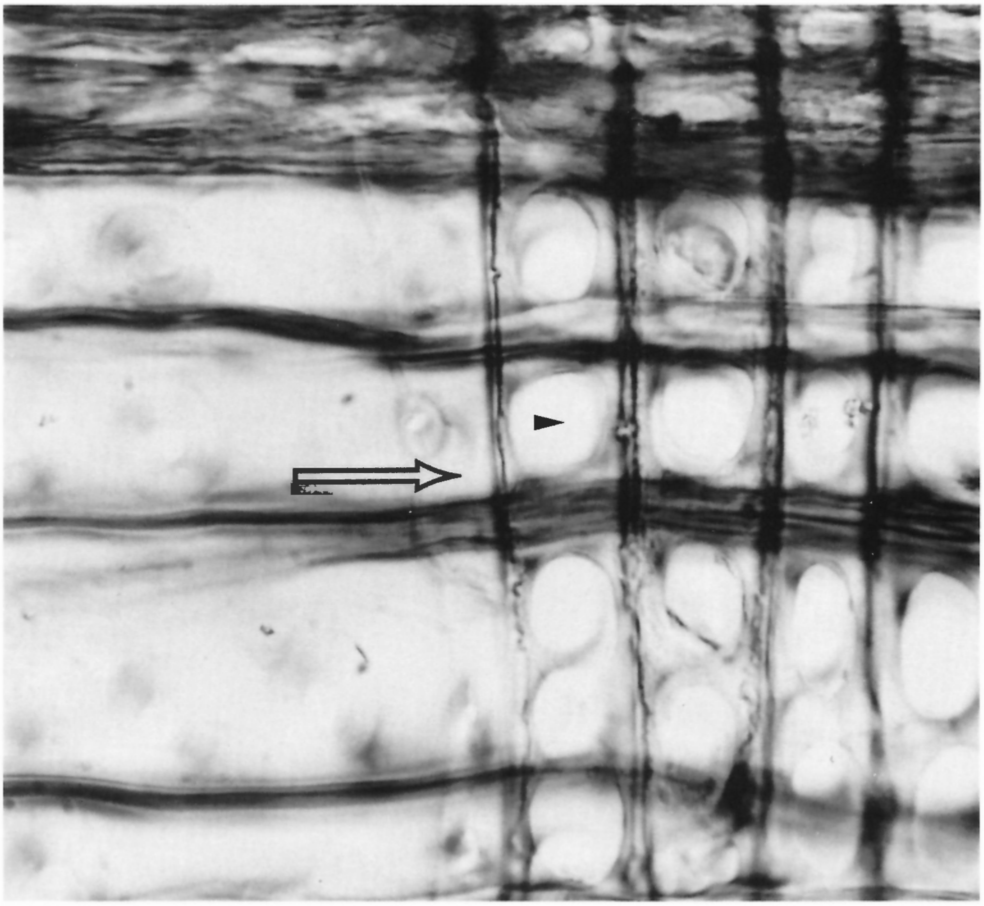

Figure 2. *Light micrograph of a radial section of the Miocene fossil wood Resolute from Cornwallis Island. Characteristics that mark this specimen as white pine are nondentate ray tracheid (arrow) and window-like (fenestriform) cross-field pit (arrowhead). Magnification x510.*

substance, as shown by subsequent analyses reported in this paper.) These deposits were not evident in cells that appeared normal. In specimens of 500 000-year-old spruce, the electron-dense material coats the surfaces of the tracheids (Fig. 4C, asterisks). The middle lamella region between cells and at cell corners is intact, but most structural integrity has been lost. The remaining secondary wall layers vary in size within a cell (Fig. 4C). The electron-dense deposits also vary in thickness.

Examination of the Miocene fossil wood Resolute showed that cells are degraded to different extents (Fig. 4D). The secondary wall layers are still evident in some cells but completely removed in other cells, leaving only the middle lamella. Different amounts of electron-dense material coat the tracheids as well as the ray parenchyma cells (Fig. 4D).

Sectioned Eocene fossil woods were also inspected. The latewood region of specimen GDH 28 shows tracheids with large secondary walls that are in the process of alteration. Cracks filled with electron-dense material were seen within the S_2 layer (Fig. 5A). The cells appear to be in a state of transition, changing from clearly discernible cell wall substance to the electron-dense deposits. The middle lamella and the S_1 layer remain, but they have little integrity. The bending of cell walls is most prevalent at the midpoint in the middle lamella between cell corners, as observed in sections of specimen GDH 61 (Fig. 5B). In this specimen, the electron-dense deposits are extensive and have two distinct layers, an inner layer adhering to the cell wall (Fig. 5B, asterisks) and an extremely electron-dense outer layer exposed to the cell lumina (Fig. 5,B, arrowheads). The remaining cell wall structure, although extensively altered, remains clearly visible and distinct from the electron-dense areas.

The cells of specimen GDH 80 (Fig. 5C) are similar in appearance to those of GDH 61 shown in Figure 5B. A two-layered electron-dense area is present in all cells. Some cells, however, show compression wood, with thick secondary wall layers and voids located at cell corners. The compression wood cells seem to retain greater structural integrity than the other cells.

Of the woods examined, the Paleocene specimen HWC 37 has the most degraded cells (Fig. 5D). Individual cells are not easily discernible because the orientation of cells has changed drastically. In some areas of the wood, cell walls are partly

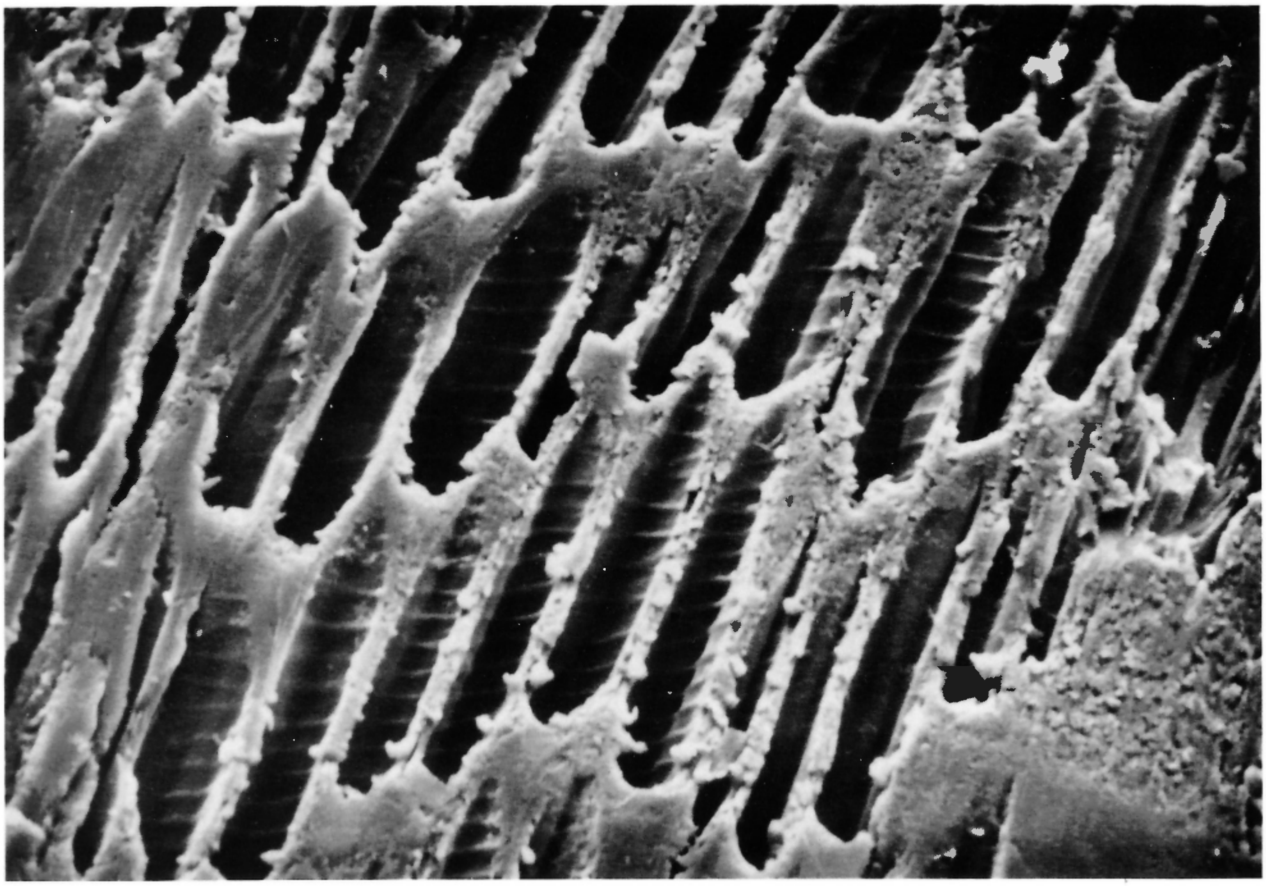

Figure 3. *Scanning electron micrograph of the Eocene fossil wood GDH 61 from Axel Heiberg Island showing spiral thickenings in the tracheids. This characteristic strongly suggests that this specimen is Douglas fir.*

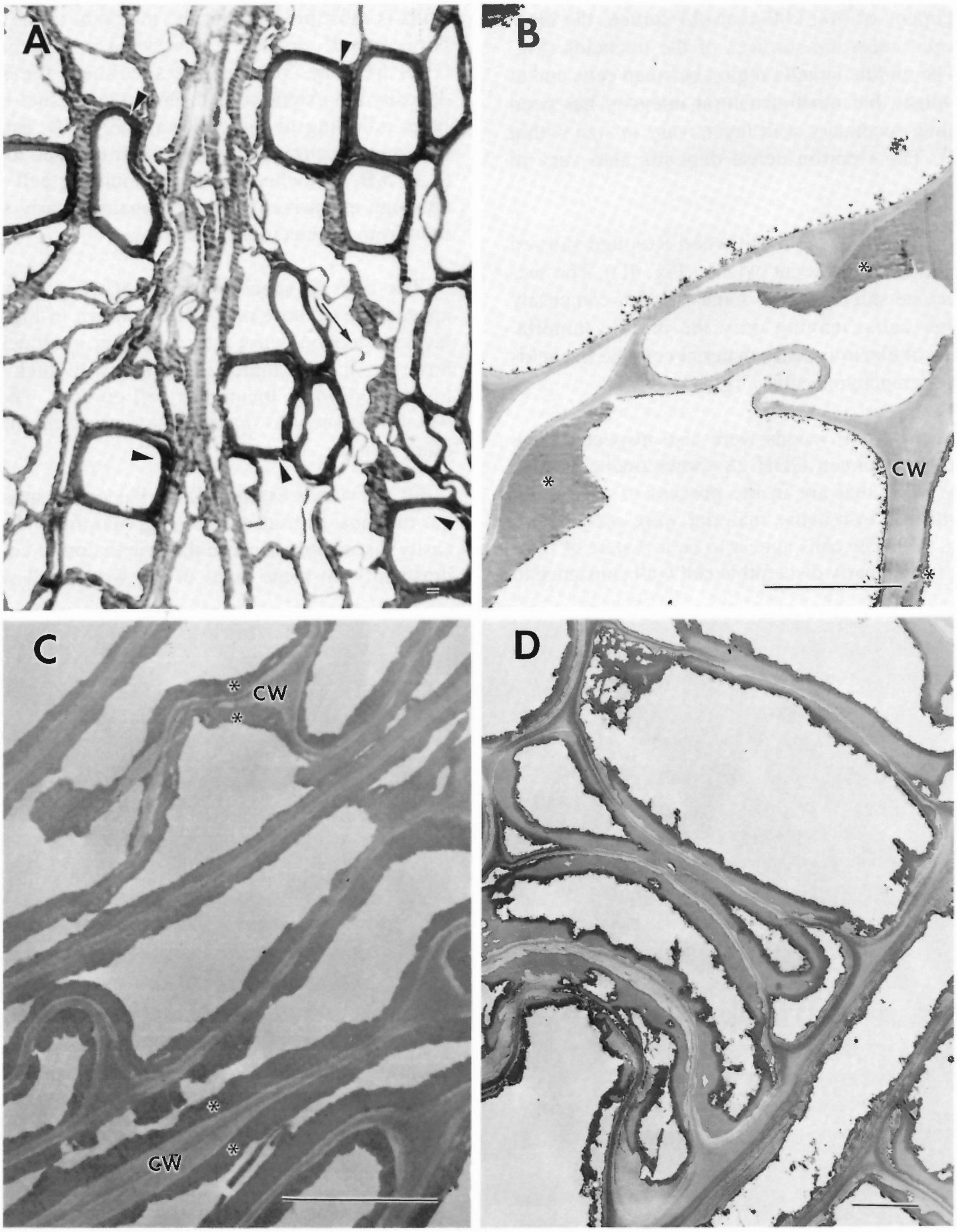

Figure 4. *Transverse sections of wood stained with Safranin and photographed using light microscopy (A) or fixed with* $KMnO_4$ *and photographed using transmission electron microscopy (B-D). A, 9900-year-old Michigan spruce showing apparently sound cells (arrowheads) and adjacent degraded cells (arrow). B, degraded cells, consisting primarily of compound middle lamella, from the 9900-year-old spruce. The secondary wall layers are no longer present in these degraded cells, whereas adjacent cells have intact cell walls (CW). Parts of the degraded cell wall have electron-dense material (asterisks). C, cells of 500 000-year-old Wisconsin spruce showing cell walls in various stages of degradation. All cells have an electron-dense material coating the walls (asterisks). D, arctic fossil wood Resolute (white pine) exhibiting tracheids with walls consisting of middle lamella and varying amounts of secondary wall layers as well as electron-dense material on cell wall surfaces. (Bar is 5 µm.)*

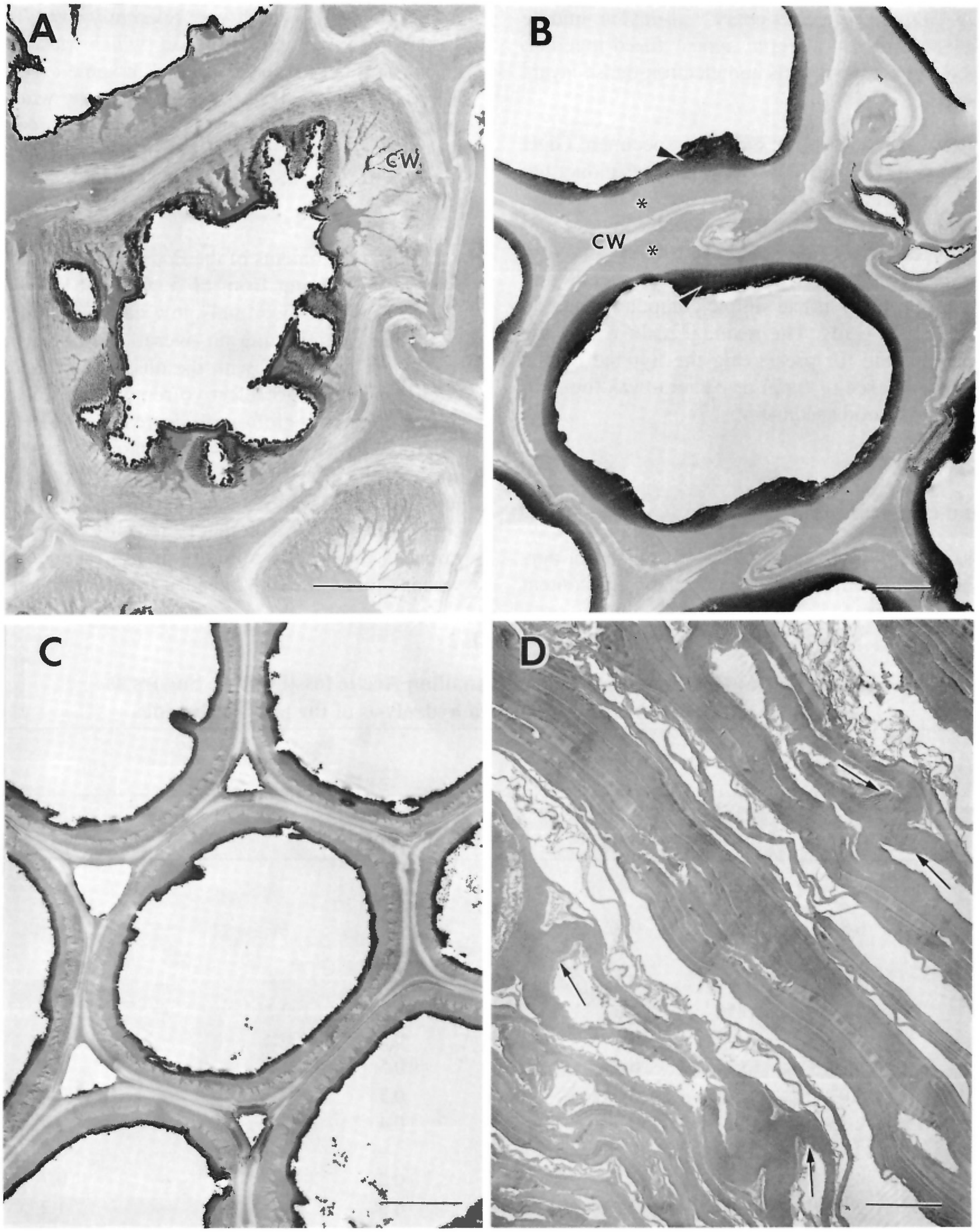

Figure 5. *Transverse sections of Arctic fossil woods fixed with* $KMnO_4$ *and photographed using transmission electron microscopy. A, latewood cells of GDH 28 showing degraded cell walls (CW). The thick secondary layers are eroded along the surface exposed to the cell lumina. Cracks within the cell wall are evident; the cell wall appears to be in a stage of transition from visually discernible cell wall substance to an electron-dense material. B, cells of GDH 61 with extensively eroded cell walls. The electron-dense material that coated the cells appears to have two layers: an inner area (asterisks) and an extremely dark outer zone (arrowheads). C, specimen GDH 80 showing degraded cell walls and two layers of electron-dense material. Some cells exhibit characteristics of compression wood, with thick secondary walls and voids at the cell corner regions. D, wood sections from HWC 37 showing extensive degradation. Cell wall layers are not clearly discernible, but a few tracheids are still evident in transverse section (arrows). The cells are compressed and appear fused with surrounding electron-dense material. The orientation of cells and lack of cell wall structure indicate that the specimen has undergone considerable degradation and compression. (Bar is 5 μm.)*

identifiable (Fig. 5D, arrows). Secondary wall layers, middle lamellae, and electron-dense material appear fused together. Loosely arranged fragments of cells and electron-dense layers were also observed.

Fungal hyphae were observed in only one specimen, GDH 61. Remnants of fungal hyphae were found in a resin duct but not in any of the tracheids. The large diameter of the hyphae and location within a resin duct suggest that the fungus could be a blue stain fungus. These fungi colonize recently killed or dying conifers and grow within ray parenchyma cells and resin ducts. Blue stain fungi utilize simple compounds and do not affect lignified cell walls. The residual resin within the duct may be responsible for preserving the hyphae in this location. No other evidence of fungi or bacteria was found in any of the arctic fossil wood specimens.

CHEMICAL ANALYSES

Ash content and chemical composition

That the fossil woods remained woody and were not permineralized was established by determining their ash content (see Appendix), which was relatively low (Table 2). Some specimens were coated with soil (which was brushed off as far as possible before analysis), and in some cases, soil particles were observed to have penetrated into the wood along cracks and check lines. Some of the measured ash content is probably not a result of replacement of organic tissue by minerals, but is rather a consequence of simple contamination during burial.

The lignin contents of the Eocene and Paleocene woods are very high, ranging from 66.5 per cent to over 85 per cent (Table 2). Correspondingly low carbohydrate contents were measured, maintaining an overall materials balance. These results are consistent with the observation that buried woods commonly undergo selective removal of polysaccharides to leave lignin-enriched residues (Crook et al., 1965). Additionally, hemicellulose is apparently removed selectively, and cellulose, as indicated by glucose content, persists the longest. The lignin content of the Miocene white pine, 48.6 per cent, is higher than that of recent loblolly pine, but not as great as the lignin contents of the Paleocene and Eocene woods (Table 2). Also included for comparison are the analy-

TABLE 2

Ash, lignin, and sugar composition of Canadian Arctic fossil woods. Sugars as anhydrides were determined after acid hydrolysis of the polysaccharides

Specimen	Ash (%)	Lignin (%)		Sugar (%)[a]					Total[b] (%)
		Klason	Acid soluble	Glu	Xyl	Gal	Arab	Man	
Paleocene									
HWC 37	0.84	83.23	0.82	13.3	—	—	—	—	98.2
Eocene									
GDH 28	2.09	65.87	0.63	27.0	0.5	0.8	—	1.4	98.3
GDH 40	2.59	70.56	0.62	29.2	0.2	0.2	—	1.6	105.0
GDH 44	5.23	81.96	0.63	12.4	0.2	0.2	—	0.8	101.4
GDH 54	3.50	70.50	0.55	25.6	0.5	—	—	1.1	101.8
GDH 59	2.02	79.76	0.66	14.4	0.3	0.2	—	0.8	98.1
GDH 61	2.35	81.27	0.52	13.9	0.4	—	—	0.8	99.2
GDH 64	2.94	78.00	0.36	19.0	—	—	—	0.7	101.0
GDH 74	4.67	83.93	0.51	10.4	0.2	—	—	0.7	100.4
GDH 80	2.50	71.80	0.57	24.1	0.3	—	—	1.2	100.5
NMNS S1	1.53	85.77	0.54	10.2	—	—	—	—	98.0
Miocene									
Resolute	1.67	47.90	0.66	37.4	2.5	2.7	—	7.6	100.4
500 000-yr-old									
Wisconsin spruce	12.02	73.73	1.35	7.5	0.9	—	0.2	1.2	96.9
9900-yr-old									
Michigan spruce	1.10	34.41	0.56	42.7	6.1	2.7	1.1	13.4	102.1
Recent									
Loblolly pine	0.30	29.80	0.59	45.6	5.4	2.6	1.5	12.9	98.7

[a] Blanks indicate sugar amount of less than 0.2 per cent. Glu, glucose; Xyl, xylan; Gal, galactose; Arab, arabinose; Man, mannose.

[b] Sum of ash, lignin, and sugars.

ses of 9900- and 500 000-year-old buried spruce specimens from Michigan and Wisconsin. The older spruce has high lignin and low carbohydrate contents (Table 2) similar to the Eocene woods. The well preserved 9900-year-old spruce has apparently lost only a slight amount of carbohydrate.

The specific gravities were measured for a few of the Eocene and Paleocene fossil woods and were found to be higher than those of recent woods. For example, the specific gravities of GDH 40, GDH 54, GDH 64, GDH 80, and HWC 37 are 1.13, 0.95, 0.82, 0.67, and 1.14, respectively. The least physically degraded specimen, Resolute, has a specific gravity of 0.38, whereas specimens of recent Douglas fir and white pine have specific gravities of 0.37 and 0.43. Although the Resolute specimen has a low specific gravity, it also shows some compression. Because of volume changes due to compression, mass losses from the original wood cannot be reliably calculated.

It is apparent that carbohydrate removal increases with the length of time that the wood is buried, leaving a lignin-enriched residue. However, the amount of degradation is related to the conditions experienced by the wood and not simply to the duration of burial. Thus, the 500 000-year-old spruce is more similar in composition to the older arctic specimens than to the Miocene specimen. The actual conditions of burial experienced by these specimens cannot be deduced exactly. In any case, the survival of any of the chemical constituents of the woods for such lengths of time is remarkable.

Extractives

Wood extractives are usually determined by extraction with hot ethanol/benzene. However, because of the degraded nature of the fossil woods, it was assumed that information derived from such an extraction would be of dubious value. As an alternative, two specimens were extracted with ethyl ether (see Appendix), and neutral and acidic fractions were separated (Table 3). The acid fraction of fossil wood GDH 74 is 0.03 per cent. However, significant wood substance invariably has been lost in fossil woods, and the acid fraction on an original wood basis would be much lower. Analysis by gas chromatography showed the presence of three saturated fatty acids, C-16:0, C-18:0, and C-24:0. Four resin acids were identified. In order of abundance, these are: abietic acid, dehydroabietic acid, isopimaric acid, and neoabietic acid. Whereas the resin acid content of most conifers is about 1 per cent of the wood, the resin acid fraction of GDH 74 is 0.01 per cent. The longevity of the abietic acid is interesting in that this acid is relatively reactive. Aromatization to form dehydroabietic acid is likely, and this compound would be expected to be present if any resin acids survived. We anticipated that decarboxylation of resin acids would yield relatively stable neutral diterpenes. However, although a small amount of dehydroabietane was identified, the neutral fraction is slight (0.01%) and almost completely composed of unknown substances.

The ethyl ether extract of the Resolute fossil wood contained even less material than did the GDH 74 extract. Although some fatty acids were identified (Table 3), no resin acids were found. The neutral fraction is only 0.001 per cent of the fossil wood and contains no identifiable components.

Acetyl, nitrogen, methoxyl, and alkaline nitrobenzene analyses

Typical softwoods have acetyl contents that range from 0.5 per cent to almost 2 per cent. Since these groups occur as substituents on the hemicellulose galactoglucomannan, no or very low acetyl would be expected to be found in the fossil woods. Selected specimens were chosen for analysis; the measured acetyl content was 0.1 to 0.2 per cent (Table 4). The low acetyl content found generally correlates to low hemicellulose, with the exception of the Michigan spruce. If acetyl groups were removed by either hydrolysis or microbial degradation, the persistence of even small amounts of residual acetyl in the much older specimens was notable.

The nitrogen content of most woods is about 0.1 per cent, a reflection of low protein content. The nitrogen content of selected fossil wood specimens shows residual values of 0.10 to 0.23 per cent (Table 4). We did not determine whether the nitrogen was present as a remnant of the wood protein or if it possibly had a nonwood origin. If the nitrogen were from the original wood protein, its persistence suggests a degree of

TABLE 3

Amount and composition of acidic and neutral fractions from ethyl ether extracts of Resolute and GDH 74 fossil woods

Extracted fraction	Amount of acid(%)[a]	
	GDH 74	Resolute
Acid fraction		
(Total weight)	(0.033)[b]	(0.020)[b]
C-16:0 fatty acid	4.3	17.5
C-18:0 fatty acid	8.7	38.5
C-22:0 fatty acid	0.0	6.0
C-24:0 fatty acid	6.6	0
Abietic acid	24.1	0
Dehydroabietic acid	15.2	0
Isopimaric acid	3.3	0
Neoabietic acid	1.2	0
Unknown	36.6	38.0
Neutral fraction		
(Total weight)	(0.011)[b]	(0.001)[b]
Dehydroabietane	5.8	0
Unknown	94.2	100.0

[a] Percentage of fraction.

[b] Percentage of total weight of wood.

stability that might result from covalent bonding to the original lignin.

Softwood lignins are composed nearly exclusively of guaiacylpropane units (Obst and Landucci, 1986) and usually have a methoxyl content of about 15.5 per cent. Methoxyl analyses were performed on the Klason lignins prepared from all the buried wood specimens; the results are given in Table 4. All the fossil wood lignins had a lower methoxyl content than recent lignin. The Paleocene specimen had the lowest methoxyl content; methoxyl loss correlated roughly with the age of the specimen. The lack of better agreement is likely due to the different aging conditions experienced by the various specimens.

Table 4 also shows the results of alkaline nitrobenzene oxidation. All the specimens produced vanillin; syringaldehyde was not detected. The yields are reported on the basis of weight percentage yield of vanillin, based on the weight of lignin in the specimen. There was a correlation between increasing specimen age and decreasing vanillin yield. As expected, there was some agreement between low methoxyl content and low vanillin yield. However, methoxyl loss itself apparently cannot explain the lower yield of vanillin in all cases. Selective degradation of noncondensed units would result in lower vanillin yields from the residual lignin. Sidechain degradation and aryl condensation reactions may have occurred during the long burial period and could also contribute to lower vanillin yields.

TABLE 4

Acetyl, nitrogen, and methoxyl analyses, and alkaline nitrobenzene oxidation of Canadian Arctic fossil woods

Specimen	Acetyl (%)	Nitrogen (%)	Methoxyl on Klason lignin (%)	Vanillin yield on nitrobenzene oxidation (weight %)[a]
Paleocene				
HWC 37	0.2	0.23	9.6	6.6
Eocene				
GDH 28	ND[b]	ND	12.1	9.9
GDH 40	ND	ND	10.0	8.5
GDH 44	ND	ND	13.1	12.1
GHD 54	ND	ND	13.4	11.7
GDH 59	ND	ND	12.4	13.2
GDH 61	ND	ND	13.4	13.7
GDH 64	ND	ND	11.9	10.9
GDH 74	ND	ND	13.1	11.7
GDH 80	0.2	0.14	11.6	11.1
NMNS S1	ND	ND	12.8	12.2
Miocene				
Resolute	0.1	0.10	14.0	18.3
500 000-yr-old Wisconsin spruce	ND	ND	12.0	12.2
9900-yr-old Michigan spruce	0.2	0.10	14.6	23.3
Recent				
Loblolly pine	1.1	0.20	15.4	25.7

[a] Based on lignin content.
[b] ND, not determined.

SPECTROSCOPY AND SPECTROMETRY

To describe the arctic fossil woods more fully, Fourier transform infrared (FTIR) spectroscopy, ultraviolet (UV) spectroscopy, solid-state and solution-state ^{13}C-NMR spectroscopy, and analytical pyrolysis/gas chromatography-mass spectrometry were performed on selected specimens (see Appendix). These methods are particularly useful for lignin characterization, which is appropriate for the fossil woods because they are composed mainly of lignin. Additionally, milled wood lignins (MWL) were prepared from three fossil woods and compared to recent wood MWLs.

FTIR spectroscopy

Infrared (IR) spectroscopy has often been used to examine isolated lignins and wood, and the method has been used to study the effect of aging on lignin (Borgin et al., 1975). Advances in FTIR, using certain new techniques, make this an important characterization tool (Faix and Beinhoff, 1988).

The normalized FTIR spectra of six fossil woods are compared to the spectrum of recent loblolly pine wood in Figure 6. All spectra display the characteristic guaiacyl band at 1267 to 1269cm^{-1} and the aromatic band at 1510 to 1512cm^{-1}, instead of at 1506cm^{-1}, which is typical of angiosperm lignins. Systematic differences can be observed: the intensities of carbohydrate C-O-C and C-O stretching (1058-1157cm^{-1}) decrease with age (from the top to the bottom of the figure). The lignin character of the woods correspondingly increases in the 1510 to 1030cm^{-1} region, and the spectrum of the Paleocene wood HWC 37 is very similar to that of a pure lignin. These observations support a progressive polysaccharide degradation of the fossil woods. Additionally, the carbonyl and carboxyl content of the residual carbohydrates is higher when the degradation is more advanced.

To illustrate these qualitative descriptions, the absorbances of six relevant bands are given in Table 5. Most absorbance values correlate with the total lignin content. Especially good correlation is apparent when the absorbance ratios 1059/1510 and 1059/1464 are submitted to a simple regression analysis with the lignin content (Fig. 7). A recent study demonstrated that lignin contents can be determined by FTIR (Faix, 1988). The data in Figure 7 suggest that the lignin contents of highly degraded gymnosperm woods can also be reliably estimated by FTIR, even if the lignin contents range to over 85 per cent.

The IR spectra of three fossil wood MWLs are compared to MWLs from *Metasequoia glyptostroboides* and loblolly pine (Fig. 8). *Metasequoia* MWL is shown to be a guaiacyl lignin with low carbonyl content. The IR intensities of this specimen at 1719, 1662, and 1595cm^{-1} are low in comparison to those of loblolly pine. The spectra of the fossil specimens and those of the recent woods are similar, but not identical. The fossil MWLs show an increase in carbonyl absorption. Also, the relative intensities of the aromatic skeletal vibration at 1422cm^{-1} are lower in the spectra of the fossil MWLs.

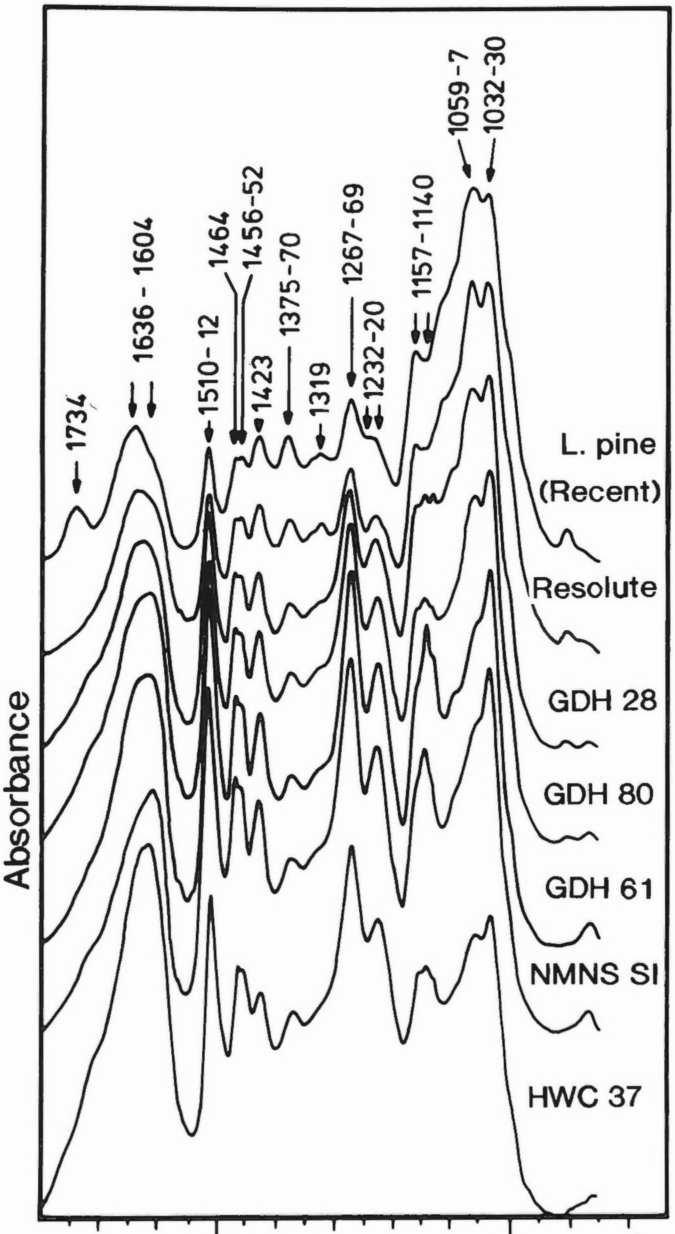

Figure 6. *FTIR spectra of recent loblolly pine and six fossil woods. These data support the conclusion from chemical analyses that the fossil woods are composed mainly of lignin.*

These data are quantified in Table 6, which compares three carbonyl band intensities. The absorbances were measured using a baseline technique (Fig. 8), and the absorbance relative to the aromatic absorbance at 1510cm^{-1} is reported (Table 6). By this method, the Resolute MWL is shown to have about the same unconjugated carbonyl content as the loblolly pine MWL, whereas the unconjugated carbonyl contents of GDH 80 and NMNS S1 MWLs are greater. These results are consistent with the degree of degradation as indicated by lignin and carbohydrate compositions. The GDH 80 MWL has the highest content of conjugated carbonyl and may be considered the most degraded of the MWLs examined. The exact meaning of the decrease in intensity of the 1422cm^{-1} band (Table 6, Aryl) in the fossil MWLs is not yet completely understood. However, this is probably a result of a greater degree of condensation. This suggestion is supported by the fact that angiosperm lignins, which have fewer sites for condensed linkages, display higher intensities in this region than gymnosperms.

Ultraviolet spectroscopy

The UV spectra of MWLs isolated from fossil woods Resolute, GDH-80, and NMNS S1 have absorptivities at 280 nm of 21.7, 24.1, and 21.7 L·g^{-1}cm^{-1}, respectively. Two MWLs were also prepared from recent loblolly pine. One MWL was not purified and has a residual carbohydrate content of about 10 per cent; the other was fractionated to yield a lignin with less than 0.5 per cent carbohydrate. The absorptivity of the crude loblolly MWL, corrected for carbohydrate, is 20.3 L·g^{-1}cm^{-1}, whereas that of the low carbohydrate MWL is 22.1 L·g^{-1}cm^{-1}. These values are typical of those found for softwood MWLs, and suggest that the fossil MWLs appear to be generally free of nonlignin contaminants.

Solid-state ^{13}C-NMR spectroscopy

The advantages of solid-state ^{13}C-NMR spectroscopy for lignin characterization are that the entire specimen is examined and there is no possibility of erroneous conclusions as a result of sample fractionation. Figure 9 compares the solid-state ^{13}C-NMR spectra of Douglas fir wood and Miocene, Eocene, and Paleocene fossil woods. The peaks of greatest intensity in the spectrum of Douglas fir wood are at 73, 76, and 106 ppm, which are assigned to polysaccharides. Lignin aryl carbons contribute a band of peaks from about 110 to 154 ppm, and associated methoxyl carbons occur at about 56 ppm. A comparison of the relative intensities of the polysaccharide and lignin peaks clearly shows a loss of carbohydrates over time. Not unexpectedly, the carbohydrates that persist appear to be mainly crystalline cellulose. This is shown by the signal at 89 ppm, assigned to C-4 of the glucose units. A signal at 84 ppm could be assigned to C-4 in disordered cellulose, to hemicellulose, or to lignin sidechains. This ambiguity was removed by exploiting differences in proton rotating-frame relaxation rates (Newman, 1987). The spectrum of the Eocene specimen GDH 80 was separated into subspectra of domains containing cellulose (Fig. 10A) and other material (Fig. 10B), with faster relaxation in the latter domains indicating greater freedom for molecular motion. The subspectrum of the cellulose (Fig. 10A) shows contributions from both crystalline and disordered material. The other subspectrum (Fig. 10B) resembles typical spectra of modern softwood lignins, except that a band from about 10 to 50 ppm is stronger and the ratio of areas assigned to sidechain and aryl carbon is lower. Relative areas provide an estimated maximum of 1.9 sidechain carbons per aryl ring, after allowance for methoxyl and weak spinning-sideband signals that overlap the signals from sidechain carbon.

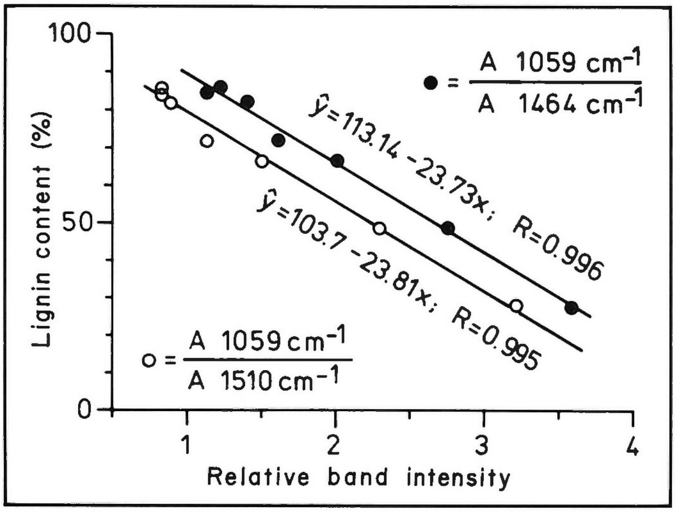

***Figure 7.** Regression equations for lignin content determinations of fossil woods from FTIR spectroscopic data.*

Figure 9 shows a band of signals from 10 to 50 ppm in spectra of the fossil woods, which becomes more conspicuous for the oldest wood. These signals could be due to resins, modified lignin sidechains, or some unidentified material. The

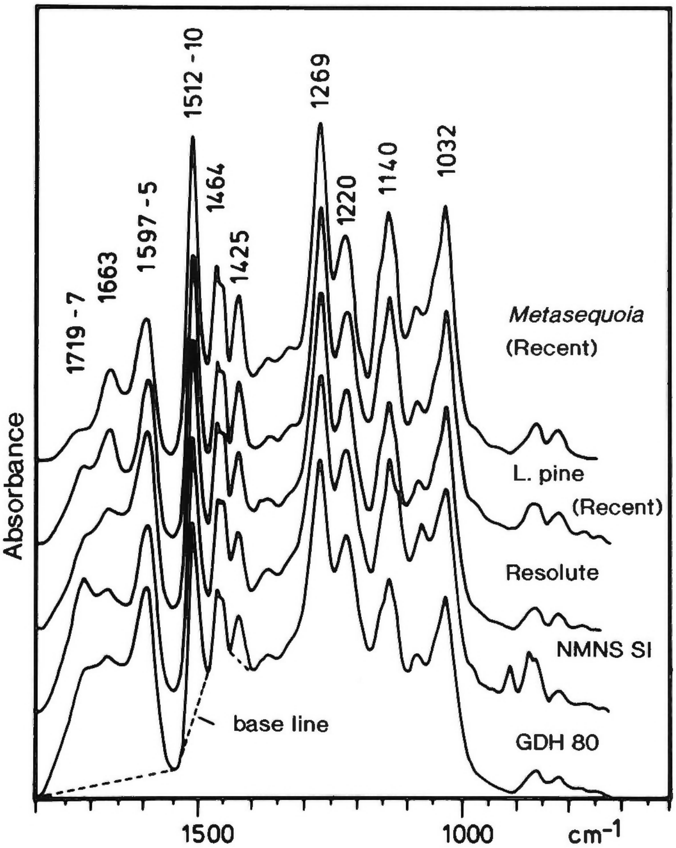

***Figure 8.** Comparison of FTIR spectra of milled wood lignins (MWLs) from recent and fossil woods.*

TABLE 5

Lignin content and selected infrared spectroscopy data for recent loblolly pine and fossil woods

Specimen	Lignin (%)	Absorbance						Relative absorbance	
		1735[a]	1635	1510	1464	1059	1030	1059/ 1510	1059/ 1464
HWC 37	84.1	0.30	1.00	0.87	0.67	0.76	0.81	0.87	1.13
GDH 28	66.5	0.13	0.55	0.63	0.48	0.96	1.00	1.52	2.01
GDH 61	81.8	0.14	0.72	0.95	0.61	0.86	1.00	0.91	1.41
GDH 80	72.4	0.15	0.66	0.79	0.58	0.92	1.00	1.16	1.59
NMNS S1	86.3	0.13	0.65	0.95	0.68	0.83	0.97	0.87	1.22
Resolute	48.6	0.07	0.44	0.43	0.36	1.0	1.00	2.30	2.74
Loblolly pine	30.4	0.15	0.36	0.31	0.28	1.0	0.98	3.22	3.57

[a] Wave numbers (cm^{-1}).

lower methoxyl contents for the fossil wood lignins reported in Table 4 could be a result of nonlignin (and nonmethoxylated) contaminants carried over from the isolation of the Klason lignins. However, close inspection of the spectra in Figure 9 shows that the methoxyl peak weakens relative to the aryl bands upon aging. This indicates at least some true methoxyl loss.

Signals from nonprotonated carbon were selected by interrupted-decoupling experiments; an example is shown in Figure 11. These spectra show broad bands and were assigned to carbonyl groups in carboxylic acids (170 to 180 ppm) and ketones (190 to 210 ppm). Peak areas indicate about 8 carbonyl groups per 100 aryl units in the Resolute, GDH 28, and GDH 80 woods, with a roughly even distribution between carboxylic and ketonic functional groups. These levels are no higher than might be expected for modern woods. Some oxidation has occurred in HWC 37, with 6 carboxylic and 13 ketonic functional groups per 100 aryl units.

Digital resolution enhancement revealed details across the band of O-substituted aryl signals from 140 to 160 ppm (Fig. 12). These signals are assigned to C-3 and C-4 of guaiacyl units, some of which are etherified through the oxygen on C-4. The degree of etherification can be estimated from the relative strengths of signals at 153 ppm (C-3 in etherified units) and 146 ppm (C-4 in nonetherified units). This NMR method was used previously to report estimates of 70 per cent etherification for Douglas fir (Leary et al., 1986) and 80 per cent for *Pinus radiata* (Leary et al., 1988). The fossil woods show some hydrolysis of ether linkages, with about 60 per cent etherification remaining in Resolute and GDH 80 woods and about 50 per cent etherification in HWC 37. The spectrum

TABLE 6

Selected IR data of MWLs isolated from recent and fossil woods

Specimen	Relative absorbance[a]			
	C = 0 Unconjugated[b]	C = 0 Conjugated[c]	C = 0 Stretch + aryl[d]	Aryl[e]
Fossil MWL				
GDH 80	0.61	0.64	0.95	0.22
NMNS S1	0.64	0.56	0.86	0.20
Resolute	0.39	0.49	0.83	0.23
Loblolly pine[f]	0.31	0.47	0.66	0.27
M. glyptostroboides[f]	0.10	0.31	0.50	0.28

[a] Absorbance relative to that at 1510cm^{-1}. Because the 1510cm^{-1} band is not significantly influenced by structural changes, quantitative comparisons may be made by comparing absorbances relative to this band.

[b] Unconjugated carbonyl at 1719cm^{-1}.

[c] Conjugated carbonyl at 1662cm^{-1}

[d] Carbonyl stretching at 1596cm^{-1} plus aromatic skeletal vibration.

[e] Aromatic skeletal vibration at 1422cm^{-1}.

[f] Recent wood.

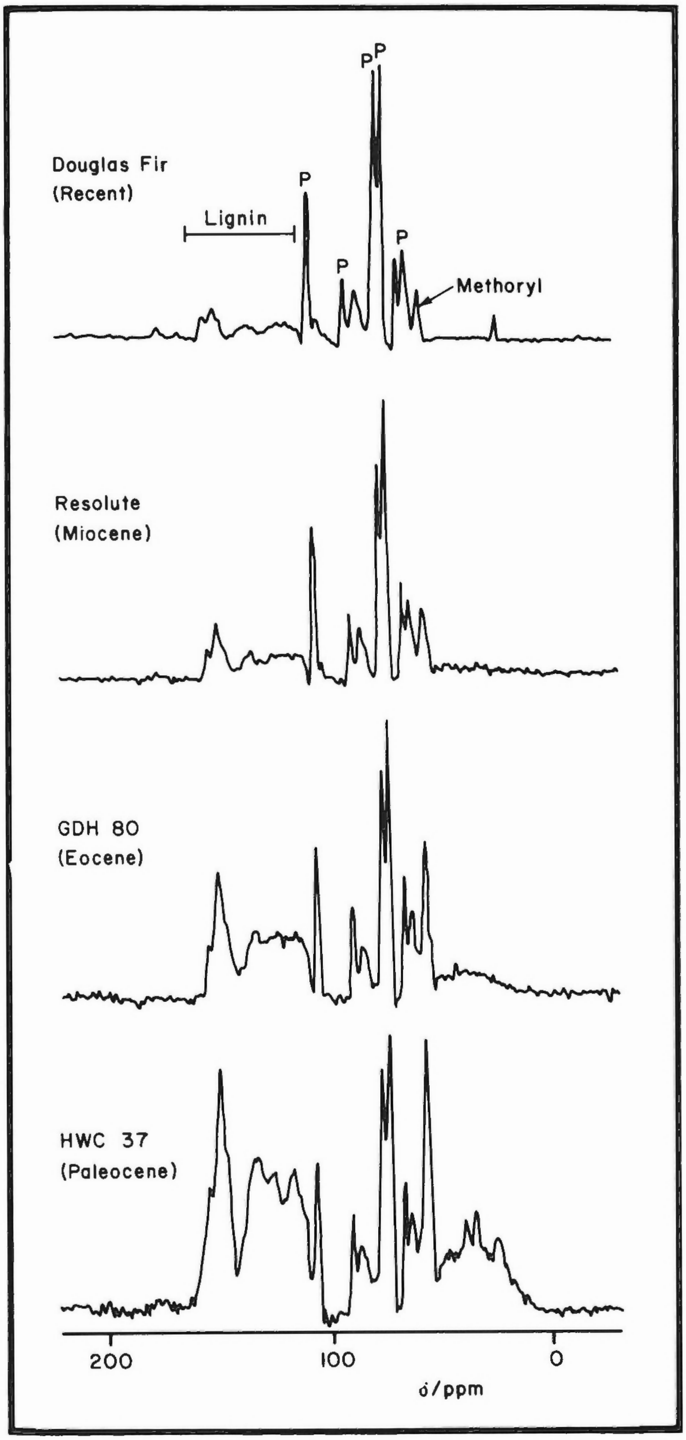

Figure 9. *Solid-state ^{13}C-NMR resolution-enhanced spectra of recent and fossil woods. P denotes major polysaccharide carbons. Most aromatic lignin carbons are in the region of 100 to 165 ppm.*

of HWC 37 shows an extra peak at 156 ppm (Fig. 12). This peak is well above the level of noise for this experiment, and it is assigned to O-substituted aryl carbon in structures with O-substitution at only one ring position. The signal strength provides an estimate of 15 per cent of aryl rings O-substituted at only one position. This is consistent with the observation of a weakened signal from methoxyl carbon, and it points to a loss of aryl oxygen. Similar effects from a loss of aryl oxygen were observed in a series of coalified gymnosperm woods (Hatcher, 1988), although only a shift in band maximum was reported rather than resolution of discrete NMR signals.

Solution ^{13}C NMR-spectroscopy

Milled wood lignins isolated from the Resolute, NMNS S1, and GDH 80 fossil woods were acetylated with acetic anhydride/pyridine, dissolved in acetone-d_6, and examined by ^{13}C-NMR spectroscopy. A quantitative method, similar to that reported previously (Landucci, 1985), was used to compare the gross features of these MWLs to those of a recent loblolly pine MWL.

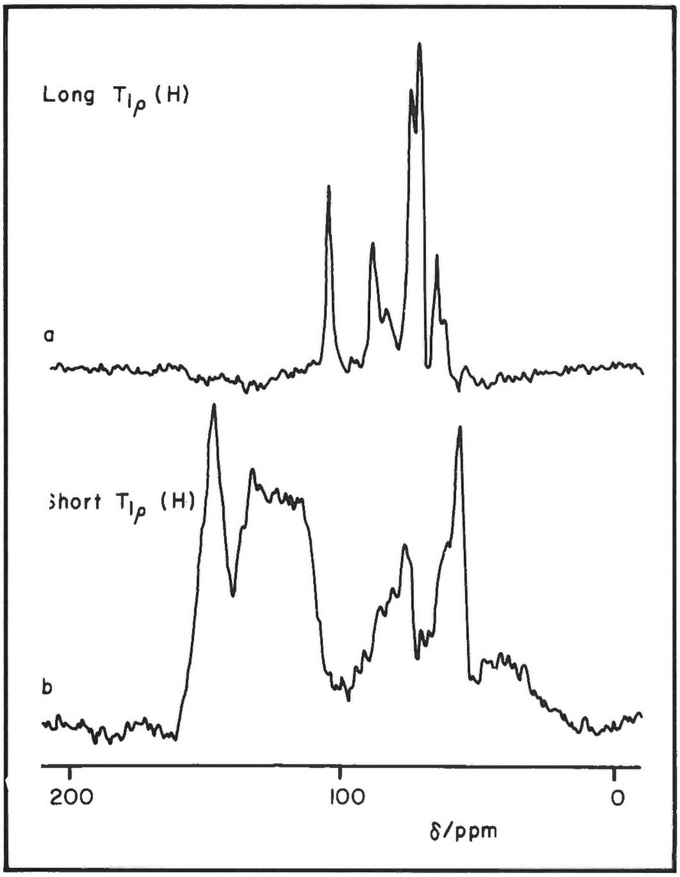

Figure 10. Solid-state ^{13}C-NMR subspectra from the Eocene wood GDH 80: A, slow relaxation component and B, fast relaxation component.

Two approaches were tried for quantitative comparisons of spectra of the four MWLs (Fig. 13). The first approach was based on the assumption that the basic lignin monomer is a C-9 phenylpropane unit (methoxyls not included). We also assumed that the signals arising from these carbons occurred in the 35 to 55.6 and 60 to 165 ppm chemical shift regions. This approach does not infer any particular aromatic/aliphatic

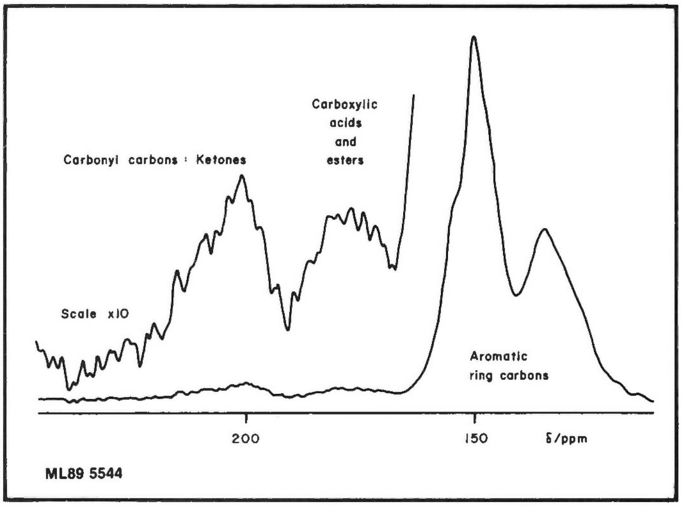

Figure 11. Solid-state ^{13}C-NMR spectrum, with interrupted decoupling, of HWC 37. Broad bands caused by carbonyl carbons in carboxylic acids and ketones are more easily seen in the expanded spectrum.

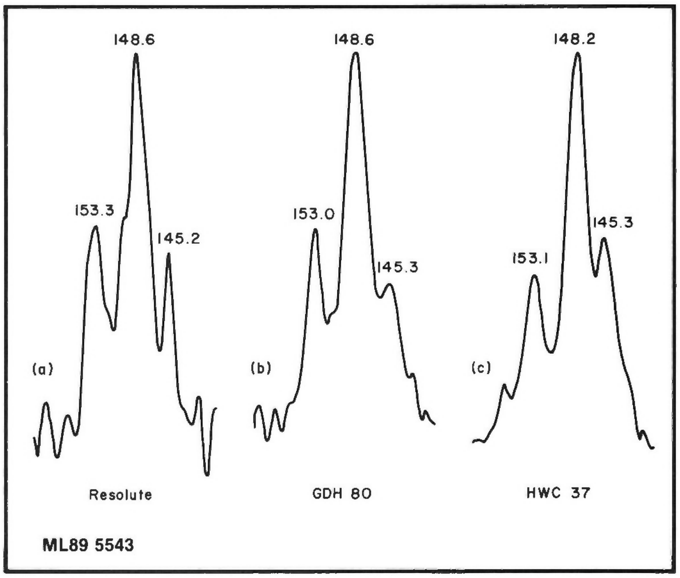

Figure 12. Portions of solid-state ^{13}C-NMR spectra run with interrupted decoupling and plotted with resolution enhancement. Signals at 153 ppm indicate etherified units (C-3). The appearance of signals at 156 ppm in the spectrum of HWC 37 may be due to O-substituted aryl carbon in structures with O-substitution at only one ring position.

ratio and is unaffected by sidechain unsaturation. The second approach assumed no sidechain unsaturation, exactly six aromatic carbons per lignin monomer, and the occurrence of these aromatic carbon signals only in the 100 to 165 ppm range. The advantage of the second approach is that it is more valid for sidechain-degraded lignins, i.e., for lignins where complete survival of C-9 units is unrealistic.

Applying these two approaches to recent loblolly pine MWL gave excellent agreement between the functionalities determined (Table 7). Because the pine MWL was not expected to contain significant sidechain unsaturation or deviate much from a C-9 polymer, this agreement was expected.

The first approach, assuming a C-9 structure, is generally less restrictive in the sense that nothing is assumed with regard to the aliphatic/aromatic ratio. If a significant amount of sidechain unsaturation has occurred, these signals would occur in the aromatic region and would be reflected in the relative ratio of "aromatic"/aliphatic carbons. However, if significant sidechain degradation has occurred to the extent that the lignin is no longer a C-9 polymer, then this approach would be invalid and it would generate abnormally high values for all the functionalities measured. With this type of degraded lignin, the second approach would be the method of choice since no assumptions are based on the condition of the sidechain, except that it cannot contain a high degree of unsaturation. Neither approach is valid for a lignin that has both sidechain unsaturation and sidechain degradation. However, this may not have been a serious impediment in our study because conditions leading to sidechain degradation are not likely also to lead to sidechain unsaturation.

TABLE 7

^{13}C-NMR functional group determination for acetylated loblolly pine MWL

Functionality	Content per monomer unit	
	Approach 1[a]	Approach 2[b]
Methoxyl	0.93	0.93
Acetoxy methyl	1.86	1.86
Acetoxy carbonyl		
Primary	0.85	0.86
Secondary	0.61	0.61
Phenolic	0.35	0.35
Aromatic	5.97	6.00[c]
Aliphatic		
60 to 100 ppm	2.78	2.79
35 to 55 ppm	0.24	0.24
Total	3.02	3.03

[a] Assumed that the average lignin monomer was a phenylpropane unit.

[b] Assumed that only aromatic carbons occurred in the 100 to 165 ppm chemical shift range.

[c] Assigned.

The result of applying both quantitative approaches to the fossil wood acetylated MWLs is presented in Table 8. In contrast to the recent loblolly pine, the fossil lignins gave quite different values by the two methods. This may have been due to an incorrect assumption that the lignin is largely composed of C-9 units (Approach 1) or that no appreciable unsaturation in the sidechain has occurred (Approach 2). For woods that have been buried for many millions of years, one can reasonably assume that significant sidechain degradation may have occurred, resulting in loss of carbon. Degradation conditions would not likely favour a net generation of unsaturated sidechains; such structures would be relatively unstable and would react further. Thus, Approach 2 is probably the more valid method for examining fossil MWLs.

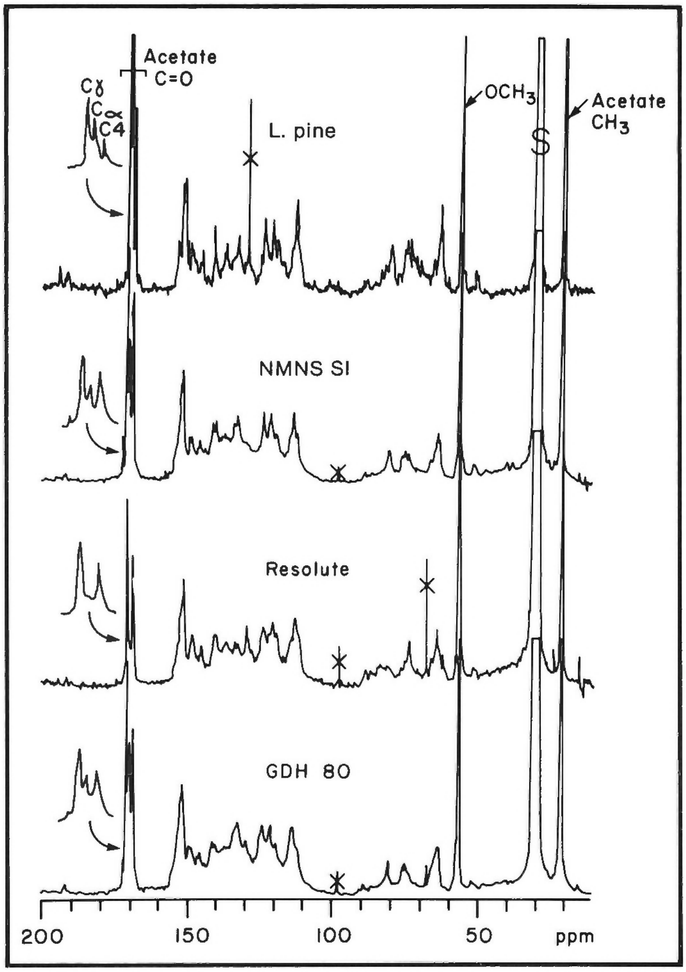

***Figure 13**. Quantitative solution-state ^{13}C-NMR spectra of loblolly pine and fossil milled wood lignins. Acetate carbonyl carbon region is shown expanded for each.*

Therefore, ^{13}C-NMR data from fossil wood MWLs are quantitatively compared to that of the recent loblolly pine MWL using the second approach. The fossil MWLs have methoxyl contents 14 to 21 per cent lower than that of the loblolly pine MWL (Tables 7, 8). These results roughly agree with the lower methoxyl contents of the Klason lignins prepared from the whole specimen (Table 4). Even more pronounced is the difference in the aliphatic acetoxy contents (from derivatization of sidechain alcohols), which are 32 to 54 per cent lower for the fossil MWLs. This decrease is consistent with sidechain degradation (removal of sidechain carbons and their associated hydroxyls). Thus, whereas the loblolly pine lignin has 3.0 sidechain carbons, the fossil lignins average <2 sidechain carbons.

The distribution of the types of acetoxy carbonyls (primary, secondary, and phenolic) is interesting. The GDH 80 and NMNS S1 MWLs have similar values but these values are very different from the Resolute MWL. The latter has an unusually low content of acetylated secondary alcohols (Fig. 13, 167 to 172 ppm expansions). The similarity of the GDH 80 and NMNS S1 MWLs is also reflected by their high acetylated phenolic contents (0.48) compared to the Resolute MWL (0.34). This high phenolic content correlates with the low methoxyl values determined for GDH 80 and NMNS S1 MWLs and is consistent with extensive demethylation.

The NMR spectra were further analyzed by dividing the aromatic signals into protonated and unprotonated carbons by means of DEPT (distortionless enhancement by polarization transfer) spectra (Fig. 14, Table 9). This method detects only protonated carbons (Doddrell et al., 1982). Values for the recent pine MWL are consistent with a small amount of ring condensation, most likely at C-5, which has been well established in the literature. The slight increase in unprotonated carbons for the GDH 80 and NMNS S1 MWLs indicates a small increase in the extent of condensation of aryl units. This is consistent with the new, but rather weak, signals in the 30 to 45 ppm region caused by nonoxygenated methylene and methine carbons. Unfortunately, the intense solvent peak overlapped these signals and accurate measurements were not possible. However, the equality of the protonated and unprotonated ring carbons in the Resolute MWL is not consistent with increased condensation. In fact, these data indicate that the Resolute MWL is less condensed than the loblolly pine MWL. One possible explanation is that demethoxylation or dehydroxylation may have occurred. The appearance of a group of signals at 128 to 130 ppm (Fig. 14) supports this explanation. A DEPT experiment showed that these signals in the Resolute MWL are protonated carbons. The only low-field aromatic signals observed from acetylated lignins have been assigned to C-2 and C-6 in p-hydroxyphenylpropane structures (Nimz et al., 1981).

Although the aromatic signals of the four MWLs showed many differences (Fig. 14), no definite conclusions may be drawn because of the lack of reliable assignments for high-resolution spectra of this type. However, the 60 to 85 ppm aliphatic region (Fig. 15) clearly shows that the erythro/threo pairs for the alpha (74.6 to 75.5 ppm), beta (80.1 to 80.6 ppm), and gamma (63.1 to 63.7 ppm) sidechain carbons are present for all the lignins except the Resolute MWL, which indicates severe sidechain degradation of the Resolute MWL.

Finally, no significant signals occurred in the 190 to 195 ppm region because of alpha-ketones and alpha- and gamma-

TABLE 8

^{13}C-NMR determined functionalities of fossil MWLs using two approaches[a]

	Fossil milled wood lignins					
Functionality	GDH 80		Resolute		NMNS S1	
	1	2	1	2	1	2
Methoxyl	0.88	0.73	0.91	0.80	0.73	0.71
Acetoxy methyl	1.52	1.27	0.96	0.85	1.37	1.20
Acetoxy carbonyl						
Primary	0.52	0.43	0.46	0.41	0.51	0.45
Secondary	0.35	0.29	0.12	0.10	0.26	0.23
Phenolic	0.58	0.48	0.38	0.34	0.55	0.48
Aromatic	7.17	6.00	6.77	6.00	6.85	6.00
Aliphatic						
60 to 100 ppm	1.50	1.25	1.84	1.62	1.63	1.43
35 to 55 ppm	0.33	0.28	0.41	0.37	0.49	0.43
Total	1.83	1.53	2.25	1.86	2.12	1.86

[a]Approach 1 assumed that the lignin monomer unit was a phenylpropane. Approach 2 assumed that all signals from 100 to 165 ppm were due to lignin aromatic carbons. Approach 1 was shown to be invalid because of sidechain degradation.

aldehydes in the fossil wood MWL spectra (Fig. 13). This result conflicts with the FTIR characterization of the MWLs, which did show increased carbonyl contents for the fossil woods. Because the MWLs were acetylated for NMR analysis, it is possible, though not very likely, that some fractionation occurred and the oxidized portion of the acetylated MWLs was lost.

In light of this anomaly, we must also emphasize that all the lignin characterizations from solution-state NMR studies pertain only to the soluble portion of the total lignin. The MWL isolated from recent wood is generally representative of the total lignin, but this may not be the case for all fossil woods. With this in mind, nitrobenzene oxidations were also performed on the isolated MWLs. The vanillin yields determined for loblolly pine, Resolute, GDH 80, and NMNS S1

TABLE 9

Relative proportion of protonated and unprotonated aromatic carbons in recent and fossil MWLs as determined by DEPT spectra

Milled wood lignin	Protonated	Unprotonated
GDH 80	2.6	3.4
NMNS S1	2.4	3.6
Resolute	3.0	3.0
Loblolly pine	2.7	3.3

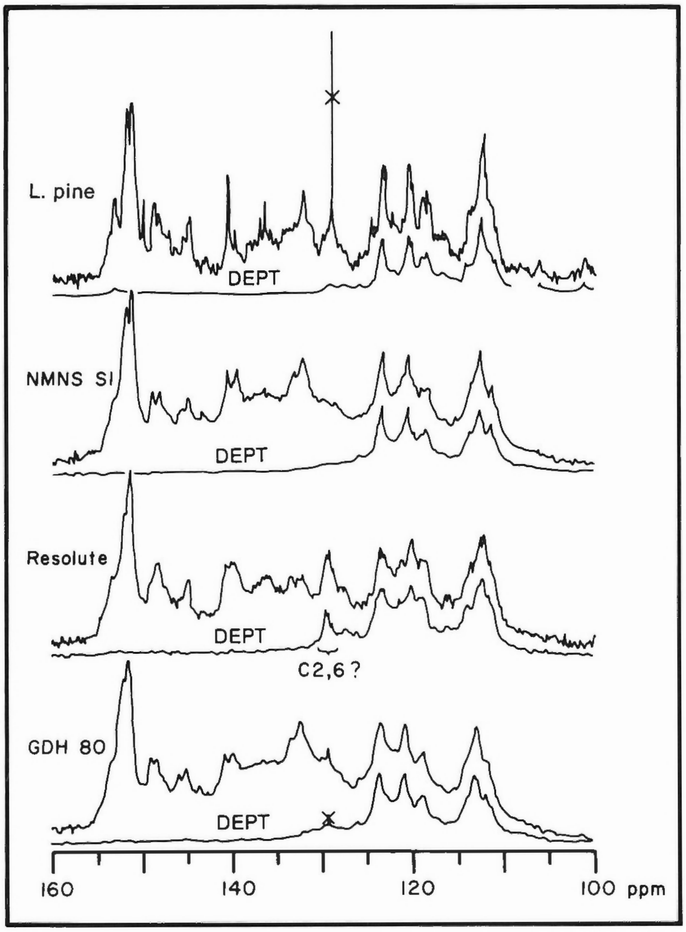

Figure 14. Partial solution-state ^{13}C-NMR spectra illustrating signals from aromatic carbons; corresponding DEPT spectra are also shown.

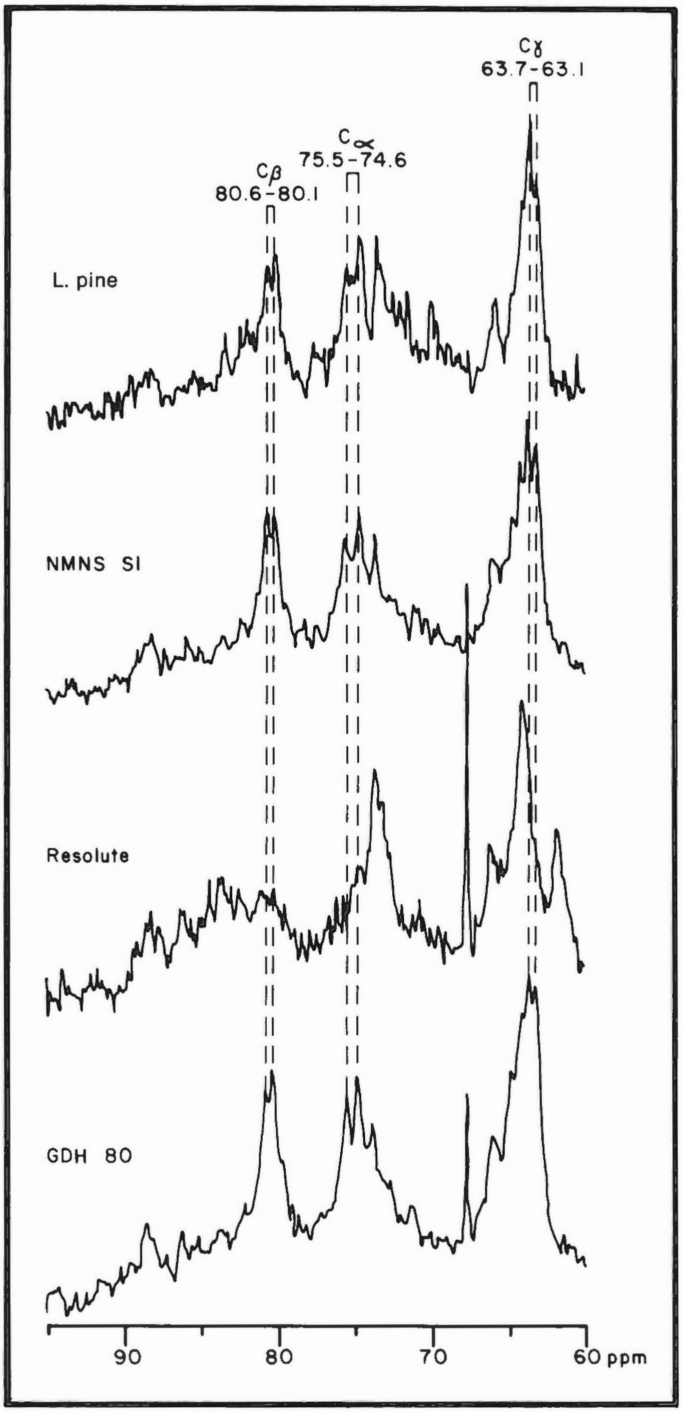

Figure 15. Partial solution-state ^{13}C-NMR spectra showing signals from oxygenated sidechain carbons. Dashed lines indicate alpha, beta and gamma erythro/threo pairs.

MWLs were 17.0, 10.7, 8.0, 8.0 weight per cent, respectively. These vanillin yields suggest that the Resolute MWL is less degraded than the other MWLs., a result again somewhat in conflict with the solution-state NMR characterization.

Analytical pyrolysis

Analytical pyrolysis (gas chromatography/mass spectrometry) has become a common technique for the analysis of lignins (Obst, 1983; Obst and Landucci, 1986; Faix et al., 1987), and this method has been applied specifically to buried woods (Saiz-Jimenez et al., 1987). Pyrograms of the Canadian Arctic fossil woods confirmed that their lignins are of the guaiacylpropane type (gymnosperm): either trace amounts of syringyl pyrolysis products or none at all were detected. However, the fossil wood and recent loblolly pine pyrograms are obviously different (Fig. 16). Generally, the fossil wood pyrograms are simpler; they contain fewer peaks caused by polysaccharide pyrolysis products and there are changes in the relative amounts of lignin-derived products.

Compared to the pyrogram of recent loblolly pine, the pyrograms of the Eocene and Paleocene woods were enriched in phenol, p-cresol, and 4-methylcatechol (Fig. 16). Lignin pyrolysis products with more complicated sidechains, such 4-vinylguaiacol, the (iso)eugenols, dihydroconiferyl alcohol, and coniferaldehyde, were less prevalent from the fossil woods. Quantitative interpretations of pyrograms is often difficult (Obst, 1983), but this arduous task has been accomplished (Faix et al., 1987). Because of the analytical data already acquired on these fossil woods, we used a simpler internal ratio method for comparing pyrograms. Selected lignin pyrolysis products were normalized to the amount of 4-methylguaiacol produced and then compared among the various specimens (Table 10). This method does not account for the yield of 4-methylguaiacol on a lignin basis, and it is useful only for comparing the product mix of the volatile components. In fact, based on the lower relative yields of vanillin from the nitrobenzene oxidations, the pyrolysis yield of 4-methylguaiacol from the fossil wood lignins is probably less than the yield from recent pine wood lignin. This relative method clearly shows that pyrolysis products with three carbon sidechains were generated in lower amounts from the fossil lignins than from the recent loblolly pine (Table 9). Relative yields of vanillin and acetovanillone from fossil wood lignin pyrolysis were generally lower than those from pine wood lignin pyrolysis.

These variances suggest significant differences in the structural characteristics of the fossil lignins compared to recent gymnosperm lignin. The MWLs isolated from the fossil woods gave pyrograms that were more similar to that of loblolly pine MWL (data not shown). This suggests that the more highly degraded lignins were not isolated on a proportionate basis. This is also reflected in the lower yields, on a lignin basis, of MWL from fossil wood compared to recent wood (see Appendix).

Compared to the loblolly pine wood, the relative amount of catechol produced from the fossil woods was the same or less.

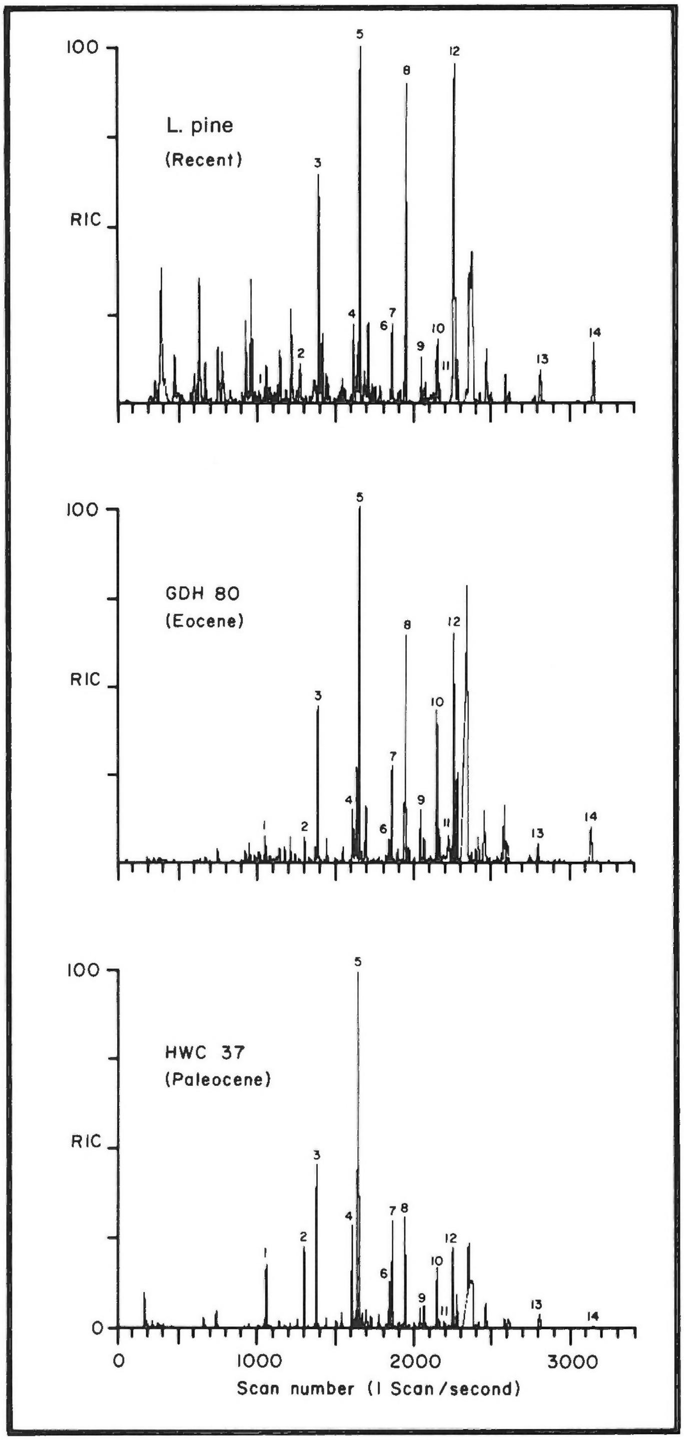

Figure 16. *Analytical pyrolysis (gas chromatography-mass spectrometry) of recent and fossil woods. For identification of some lignin pyrolysis products, the numbers refer to the compounds listed in Table 9.*

However, the relative amount of 4-methylcatechol was somewhat greater from the fossil woods compared to the recent pine. This is particularly interesting because previous research has suggested that the catechols detected in certain pyrolysis experiments are secondary products of pyrolysis resulting from the special conditions experienced using the CDS Pyroprobe[6] (see reference 18 in Faix et al., 1987). Based on a general knowledge of lignin biosynthesis and lignin characterizations, we can reasonably assume that any catechols produced by pyrolysis of normal lignins arise by demethylation of guaiacylpropane units. Although model studies have shown that the major thermal degradative pathway is demethoxylation, some demethylation may occur (Allan and Mattila, 1971). Catechol has also been identified as a wood carbonization product (Goos, 1952).

If catechol is produced from guaiacol or from units that give rise to guaiacol, then the yield of catechol from the fossil woods somewhat mirrors the guaiacol yield. The higher relative yields of 4-methylcatechol from the fossil woods might be interpreted to mean that a significant portion of this product may have arisen from demethylated (catechol) groups in the fossil lignin. The higher relative yield of catechol compared to guaiacol and the larger amount of 4-methylcatechol would especially support the observed low methoxyl content for the Paleocene specimen HWC 37.

CONCLUSIONS

Identification of the Canadian Arctic fossil woods by microscopic examination was difficult because of their poor condition. Some specimens are so degraded that they could be any of a number of conifers. However, the least degraded and most recent specimen (Resolute) was identified as white pine. Of those specimens that are in somewhat better condition, likely genera include spruce or larch. Of particular interest is one specimen that displays Douglas-fir-like spiral thickenings. The fossil woods may represent extinct species, given the difficulty of determining the species within these genera even with recent, intact specimens. Although *Metasequoia* leaves and cones have been identified from the fossil forest sites, none of the wood specimens examined in our study could be positively identified as *Metasequoia*.

Spruce, larch, and white pine generally grow in cool to temperate climates. The occurrence of these tree types suggests that the Axel Heiberg Island fossil forests may have experienced climates similar to that of present northern forests. Conifers found in present northern-most forests include white and black spruces, balsam fir, eastern hemlock, eastern pines (white, red, and jack), northern white cedar, and larch.

TABLE 10

Pyrolysis of fossil wood specimens[a]

		Product molecular ion/4-methylguaiacol molecular ion count				
Product no.[b]	Pyrolysis product	Loblolly pine	Resolute	GDH 80	NMNS S1	HWC 37
1	Phenol	0.034	0.005	0.093	0.012	0.326
2	p-Cresol	0.021	0.070	0.080	0.023	0.220
3	Guaiacol	0.632	0.479	0.406	0.317	0.405
4	Catechol	0.414	0.397	0.260	0.268	0.424
5	4-Methylguaiacol	1.0	1.0	1.0	1.0	1.0
6	4-Methylcatechol	0.053	0.133	0.077	0.121	0.148
7	4-Ethylguaiacol	0.180	0.181	0.198	0.217	0.230
8	4-Vinylguaiacol	1.039	0.681	0.729	0.373	0.327
9,10,12	Eugenols[c]	1.112	0.745	0.935	0.573	0.352
11	Vanillin	0.353	0.257	0.055	0.195	0.195
	Acetovanillone	0.152	0.144	0.272	0.102	0.107
13	Dihydroconiferyl alcohol	0.127	0.080	0.051	0.045	0.040
14	Coniferaldehyde	0.389	0.114	0.199	0.053	0.014

[a]The pyrolysis products were compared using loblolly pine as a control.
[b]Number refers to product designation in Figure 16.
[c]Eugenols are the sum of eugenol and cis- and trans-isoeugenol.

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[6] The use of trade or firm names in this paper is for reader information and does not imply endorsement by the U.S. Department of Agriculture of any product or service.

Chemical analyses showed that although the fossil woods have not been mineralized, extensive carbohydrate losses have occurred. Small amounts of resin and fatty acids could be detected. Acetyl contents are low, as anticipated by the nearly complete removal of hemicellulose from most specimens. Especially for the oldest specimens, most of the residual wood substance consists of lignin. Further characterization revealed that the surviving lignin is more condensed, demethylated, and sidechain-degraded than recent lignin.

The FTIR analysis provided a reliable measure of lignin concentration, even for fossil woods of very high lignin content. Although these characterizations tend to describe how the wood has changed, they also indicate that considerable wood structure has remained. It is also noteworthy that after tens of millions of years, cellulose and lignin have persisted in clearly recognizable forms.

These physical, chemical, and spectroscopic characterizations allow some speculation on the nature of fossil wood degradation. If the trees were rapidly buried by a flood, the conditions of aging would have been anaerobic. Thus, significant fungal decay would not have been likely to occur initially because molecular oxygen would have been essential for that process (Zeikus, 1980). Anaerobic bacterial degradation could result in removal of polysaccharides (Brenner et al., 1984), but not in lignin degradation (Zeikus, 1980). Chemical hydrolysis could occur, especially if the soils were acidic, but it would be very slow. Such hydrolysis could depolymerize the polysaccharides and the lignin, but to a much more limited extent in the case of lignin.

The preferential degradation of the carbohydrates and the high residual lignin content of the Arctic fossil woods are somewhat comparable to the chemical changes that occur in wood after degradation by brown rot fungi. However, microscopic evidence showed that the degradation is not typical of any know form of bacterial or fungal decay. Distinct morphological characteristics of cell wall attack associated with white, brown, or soft rot fungi (Wilcox, 1968; Liese, 1970; Nilsson, 1974; Hale and Eaton, 1985; Otjen and Blanchette, 1986; Blanchette, 1987) and cavitation, tunneling, and other forms of bacterial degradation (Liese, 1970; Nilsson and Singh, 1984; Daniel et al., 1987) were not observed.

Micrographs of the Arctic fossil woods showed degradation of the secondary cell walls leaving a relatively intact, but weakened, compound middle lamella region. As the secondary wall is degraded, an amorphous electron-dense material accumulates on the residual cell wall. Energy dispersive X-ray microanalysis with scanning electron microscopy showed no increase in silicates or other extraneous materials (unpublished data). Apparently, this electron-dense material originated as the cell wall layers decomposed. Specimens GDH 28, GDH 80, and HWC 37 progressively demonstrate a possible sequence of how this material accumulates. In specimen HWC 37, all the cell wall layers are indistinct, compressed, and surrounded by amorphous electron-dense material. In contrast to brown-rotted wood, which is fragile and easily crushed to a powder when dry (Blanchette, 1980), the Arctic fossil wood specimens are extremely hard and brittle. The micrographs presented suggest that simple hydrolysis could have been the major route of degradation.

ACKNOWLEDGMENTS

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APPENDIX

MATERIAL AND METHODS

Specimens obtained from the Institute of Sedimentary and Petroleum Geology, Geological Survey of Canada are listed in Appendix Table 1.

Specimen collection

The Resolute specimen was collected by A.L. Washburn, University of Washington, from a road cut at Resolute on Cornwallis Island.

The Geodetic Hills specimen obtained from David Grattan of the Canadian Conservation Institute, National Museum of Natural Sciences, was collected by James Basinger, University of Saskatchewan, in 1986. It was taken from a large, excavated stump and is identified as NMNS S1 (National Museum of Natural Sciences Stump 1).

The spruce specimen from Michigan referred to in the text was excavated in Marquette County (Traver, 1980). Exact documentation of the spruce wood from Wisconsin was unavailable to us. The specimen was obtained from the collection of Robert Maeglin at the Forest Products Laboratory along with this information: "Spruce from glacial drift at Woodville, Wisconsin; collected by Mr. Weidman, State Geologist, July, 1915; estimated to be 500 000 years old."

Ultrastructural studies

For anatomical studies, nine of the Arctic wood specimens were examined using a Zeiss light microscope. Thin (unstained) sections were prepared by hand and mounted on slides using a glycerine/ethanol solution. The Resolute specimen was also prepared by cutting thin sections on a microtome, staining with Safranin, and mounting on the slide with Cover Bond. Specimens for scanning electron microscopy were placed on specimen mounts using double-sided tape, gold coated, and then viewed with a JEOL 840 scanning electron microscope.

For ultrastructural studies, small segments were cut from specimens GDH 28, GDH 61, GDH 80, HWC 37, Resolute, 500 000-year-old Wisconsin spruce, and 9900-year-old Michigan spruce and fixed in 1.0 per cent aqueous potassium permanganate for 3 h. Specimens were dehydrated and embedded in Quetol 651 epoxy resin (Abad et al., 1988). Thin sections, 80 to 120 nm, were cut with a diamond knife and examined with a Hitachi 600 transmission electron microscope. Sections of embedded specimens were stained with 1.0 per cent aqueous Safranin and examined with light microscopy.

Chemical analyses

Ash, methoxyl, nitrogen, lignin, and sugar analyses were performed at the Forest Products Laboratory by the usual methods. Nitrobenzene oxidations were as described in Kirk and Obst (1988).

For ethyl ether extractives, 60-mesh Wiley-milled wood (1 to 2 g) was weighed and placed in a glass column. The column was eluted with 250 ml of diethyl ether. After removing the solvent, the residue was fractionated into neutral and acidic components using a semimicro DEAD-Sephadex column (Zinkel and Han, 1986). A C-17:0 fatty acid was added as an internal standard to the acid fraction, which was then methylated with diazomethane. Both fractions were analyzed by fused silica capillary column chromatography using a BDS (1,4-butanediol succinate) column at 190°C (Han and Zinkel, 1986) and DB-1 (methyl silicone) column at 170°C (Han and Zinkel, 1988).

Milled wood lignins were prepared by extracting vibratory-ball-milled wood with 90 per cent dioxane (Obst and Kirk, 1988). A low-carbohydrate loblolly pine (*Pinus taeda*) MWL was prepared in about a 1 per cent yield based on lignin in the wood. The yields of MWL, on a lignin basis, from loblolly pine, Resolute, GDH 80, and NMNS S1 were 21.2, 10.2, 13.0, and 12.6 per cent, respectively.

Instrumental characterizations

The UV spectra of the MWLs were run in 90 per cent aqueous dioxane. The IR spectra were run on potassium bromide pellets, using 1 mg of sample in 300 mg of potassium bromide. A BioRad Digilab FTS40 FTIR spectrometer, with a DTGS/KBr detector, was used for collecting the interferograms. The spectra, from 64 scans, were baseline corrected and standardized.

Solid-state ^{13}C-NMR spectra were run at 50.3 MHz on a Varian XL-200 spectrometer. Specimens of 0.11 to 0.23 g of wood were packed in 7-mm sapphire rotors with Kel-F caps and spun at 5 KHz in a DSI-206 magic angle spinning probe (Doty Scientific, Inc.). Spectra were obtained using proton radiofrequency fields of 9 G for the Hartmann-Hahn match and 15 G for decoupling during data acquisition. Proton spin-lattice relaxation times were <0.1 s with a 0.3-s pulse delay for signal recovery. Each 2-ms contact was followed by 30 ms of data acquisition, and transients from between 1500 and 10^4 contacts were averaged. Some spectra were replotted with resolution enhancement, involving convolution of the raw data with an exponentially increasing curve with time constant 3 ms (Fig. 9) or 2 ms (Fig. 12) and a Gaussian curve with time constant 6 ms. Separation of subspectra was achieved with a delayed-contact pulse sequence (Newman, 1987) run with and without a proton spin-locking pulse of 6 ms. The interrupted-decoupling experiments used the sequence of Harbison et al. (1985) with an interruption interval of 42 s and up to 46 x 10^3 transients averaged.

Solution-state ^{13}C NMR-spectra were run on MWLs that had been acetylated by treatment with acetic anhydride:pyridine (1:1) overnight at room temperature. Acetone-d_6 solutions (15 to 20%) of the acetylated MWLs in 10-mm tubes were used for all NMR experiments. The spectra were obtained with a Bruker WM-250 spectrometer controlled by an Aspect 2000A minicomputer. With a 10-mm probe tuned to 62.9 MHz, 20 000 transients of 16K data points were accumulated for the quantitative spectra and 5000 for the DEPT spectra. Standard Bruker microprograms were used. Data processing and integration have been previously described (Landucci, 1985).

Analytical pyrolysis was performed using a Chemical Data Systems (CDS) Pyroprobe coupled to a Finnigan 4510 gas chromatography/mass spectrometer. The pyrolysis was done in quartz tubes using a coil probe at 700°C, with the ramp off. The interface was at 220°C and the 60-m DB-5 capillary column was programmed from 80°C (1 min) to 220°C at 4°/min. Mass spectra were obtained at 70 EV. Molecular ion peak areas were measured for selected products and were normalized to that of 4-methylguaiacol. Because the abundance of the various molecular ions was not the same for each product, the data presented in Table 10 are useful only for comparisons between specimens, and they are not a measure of the relative yield of the pyrolysis products within a pyrogram.

APPENDIX TABLE 1

Specimen numbers and description of samples

Field number	GSC no.	Samples of wood[a]
GDH 28	164581	Trunk or branch from non-rich sandstone
GDH 40	164594	Trunk or branch lying on forest floor litter layer
GDH 44	164598	Trunk or branch containing good growth rings
GDH 54	164606	Trunk or branch
GDH 59	—	Trunk or branch in grey shale
GDH 61	164610	Trunk or branch
GDH 64	164613	Trunk or branch lying on forest floor plant litter
GDH 74	164622	Trunk or branch
GDH 80	164227	Trunk or branch
HWC 37	111498	Wood from in situ stump

GDH, Geodetic Hills, Axel Heiberg Island; HWC, Hot Weather Creek, Ellesmere Island.

[a] All samples are of *in place* specimens – meaning place of burial, not place of growth.

A PRELIMINARY PETROGRAPHIC STUDY OF COALS OF THE FOSSIL FOREST SUCCESSION, EASTERN AXEL HEIBERG ISLAND, CANADIAN ARCTIC ARCHIPELAGO

F. Goodarzi[1], T. Gentzis[2], and B. Ricketts[3]

Goodarzi, F. et al., 1991. A preliminary petrographic study of coals of the fossil forest succession, eastern Axel Heiberg Island, Canadian Arctic Archipelago. In Tertiary Fossil Forests of the Geodetic Hills, Axel Heiberg Island, Arctic Archipelago, R.L. Christie and N.J. McMillan (eds.); Geological Survey of Canada, Bulletin 403, p. 147-157.

Abstract

The petrography of low-rank coals from Axel Heiberg Island in the Canadian Arctic was investigated. The coal beds are part of the fossil forest succession in the Buchanan Lake Formation, of Middle Eocene age. The coals consist mainly of huminite (with humotelinite the dominant maceral); liptinite and inertinite are minor components.

The coal is autochthonous, and probably developed in a forest environment dominated by conifers such as *Metasequoia* and spruce, as well as hardwoods. The fossil forest coal is similar to the Hat Creek deposit in British Columbia; both have a low boron content, indicating deposition in a fresh water environment.

Résumé

On a étudié la pétrographie des charbons de bas rang provenant de l'île Axel Heiberg dans l'Archipel arctique canadien. Les couches de houille font partie de la "succession de forêts fossiles" de la formation de Buchanan Lake, qui se situe dans l'Éocène moyen. Les charbons se composent principalement d'huminite (l'humotelinite étant l'élément macéral); la liptinite et l'inertinite sont les composants secondaires.

Le charbon est de caractère autochtone, et s'est probablement formé dans un milieu forestier dominé par des conifères tels que *Metasequoia* et des épinettes, et également par des feuillus. Le charbon des forêts fossiles est semblable à celui des gisements de Hat Creek en Colombie-Britannique; tous deux ont une faible teneur en bore, ce qui indique une sédimentation dans un milieu d'eau douce.

INTRODUCTION

The fossil forest succession occurs within the Middle Eocene Buchanan Lake Formation, which overlies the Iceberg Bay Formation and consists of conglomerate, sandstone, mudstone, and coal.

The area (Fig. 1) was initially mapped during Operation Franklin in 1955 and subsequently by Thorsteinsson (1971). In addition, stratigraphic and sedimentological studies have been conducted by Hills and Bustin (1976); Bustin (1982) and Ricketts and McIntyre (1986).

The purposes of this study are: 1) to determine the petrographic characteristics of the coal seams in the fossil forest and adjacent successions using both reflected (white) and fluorescent (ultraviolet) light microscopy; 2) to compare the depositional environment, as determined by petrography (Figs. 3, 4, 5), to that determined by stratigraphic and sedimentological studies (Ricketts, *this volume*); and 3) to compare the fossil forest coal to that of other low-rank deposits, such as that in Hat Creek, British Columbia.

GEOLOGY

The overall depositional setting of the fossil forest succession in eastern Axel Heiberg Island (Fig. 1) is a broad alluvial plain (Bustin, 1982) situated in front of a rising mountain belt (Fig. 21 of Ricketts, *this volume*). During Middle Eocene time, the entire eastern Arctic region underwent structural reorganization. West- to northwest-directed compression gave rise to Eurekan tectonism, thus producing foreland-style folding and faulting. For a more detailed description of the stratigraphic, sedimentological and structural setting of the fossil forest succession, the reader is referred to Ricketts (*this volume*).

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[1] Geological Survey of Canada, Institute of Sedimentary and Petroleum Geology, 3303 - 33rd Street N.W., Calgary, Alberta T2L 2A7

[2] Alberta Research Council, Coal Research Centre Devon, One Oil Patch Drive, Devon, Alberta T0C 1E0

[3] Geological Survey of Canada, Cordilleran Division, 100 West Pender Street, Vancouver, British Columbia V6B 1R8

STRATIGRAPHIC SETTING AND COAL CHARACTER

The coal seams and lenses of coaly matter range in thickness from a few centimetres to 1.5 m and are commonly interbedded with mudstone. Root structures penetrate to depths of 1.5 m and the coal contains beautifully preserved forest flora, dominated by *Metasequoia*, spruce and hardwoods (Ricketts, *this volume*). Tree stumps in prone and growth positions, as well as branches and twigs showing various degrees of compaction are also very common characteristics of the fossil forest succession (Ricketts, *this volume*). Other macroflora, such as seeds, cones and leaves are also very abundant and spectacularly preserved in the exhumed fossil forest sites. Resin is common and occurs both as irregular lumps or small tear-shaped droplets.

SEDIMENTOLOGY

Ricketts (*this volume*) recognized four distinct facies for the fossil forest site west of the hanging wall of the Stolz Thrust. They are present within the Buchanan Lake Formation and are: 1) distal alluvial fan, probably transitional to proximal braidplain; 2) proximal braidplain; 3) transitional facies from braidplain to meanderplain; and 4) meander channel sequences indicative of extensive flood basin and crevasse splay deposition.

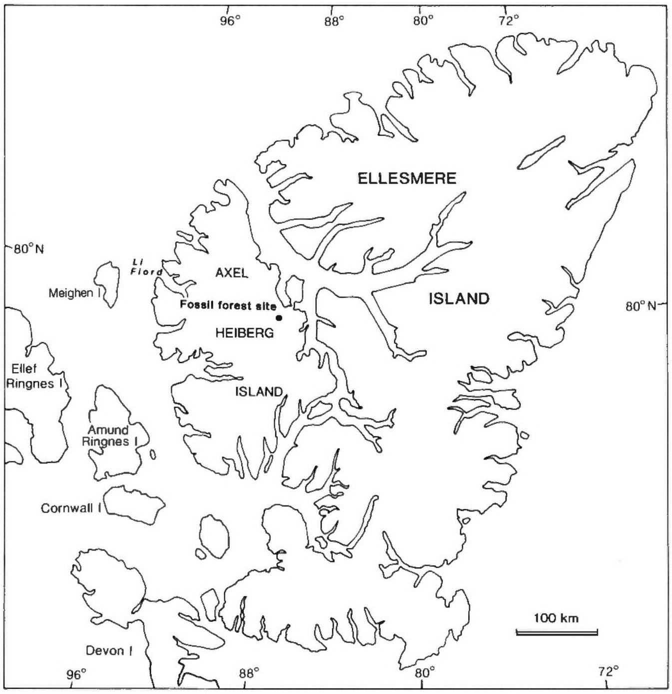

***Figure 1**. Map of Axel Heiberg Island showing location of fossil forest site.*

METHODS

For microscopic study, grain mounts were made from crushed samples of coal and interbedded sediment (-20 mesh, <850 μm), mixed with epoxy binder. Thirty samples from six sections (Table 1) were prepared and polished according to ASTM (American Standards for Testing and Materials) procedures (1981).

Reflectance measurements were carried out by means of a Zeiss MPM II microscope fitted with white (halogen) and ultraviolet light sources, using an Epiplan-Neofluar oil immersion objective (NA 0.90 x 40). Random reflectance measurements were taken on eu-ulminite B and/or (in the absence of eu-ulminite B) eu-uliminte A, using oil with *n* oil of 1.518 at 23°C. Fluorescence spectra of liptinite macerals were taken using a blue-violet excitation filter (400-440 nm), dichroic beam splitter (460 nm) and barrier filter (470 nm).

RESULTS AND DISCUSSION

Petrology

The coals (<10% by volume mineral matter content) are rich in huminite group macerals, particularly humotelinite (xylite). Humotelinite in the coal samples averages 57.0 per cent (Table 1) (Plate 1). In the interbedded sediment samples (>20% mineral matter) humotelinite averages 12.0 per cent, humocollinite 9.0 per cent and humodetrinite 3.0 per cent.

Liptinite content in the samples never exceeds 6.0 per cent (Table 1) and is dominated by sporinite, resinite, cutinite, and lesser amounts of suberinite and exsudatinite (Plate 2). The samples are also poor in inertinite (3.0%), with the exception of one interval (18.0%) (Table 1) dominated by fusinite, semifusinite and sclerotinite (Plate 1).

Humotelinite and humocollinite form from the lignin and cellulose of plant cell walls. These macerals are present in coal formed from peat that accumulated under conditions that were favourable for the preservation of woody tissue. Anaerobic conditions, generally regarded as a prerequisite for huminite formation and preservation, translate to: 1) fairly low pH of the swamp water (<4.5), which suppresses bacterial degradation; and 2) a sufficiently high water table to prevent extensive oxidation (Renton and Cecil, 1979).

Humotelinite forms in situ (is autochthonous) in areas inhabited by trees. Humodetrinite and the intervals rich in sporinite and cutinite attest to the presence of a reed marsh. This environment is inhabited by herbaceous vegetation (mainly shrubs), which produces large quantities of spores, and indicates a hyp-autochthonous setting.

The predominance of humotelinite suggests a high input of woody material (Goodarzi and Gentzis, 1987). The plant communities in the fossil forest succession consist mainly of *Metasequoia* and other conifers (spruce and pine), suggesting a forested type of swamp environment (Blackburn, 1982).

Inertinite macerals are mainly derived from the same plant components that form huminite. They do not undergo humification and gelification, but instead undergo fusinitization, which may be caused by oxidation, charring, mouldering and/or fungal attack prior to or during peat accumulation. As a result, fusinite forms in peat layers that have been subjected either to swamp fires or strong oxidation. The near absence of inertinite in the fossil forest samples indicates that the swamp deposits were rarely subaerially exposed and that the conditions favouring inertinite formation were uncommon, possibly because of the humid climate in the area.

REFLECTANCE AND FLUORESCENCE

Reflectance values of the samples studied vary from 0.14 to 0.47 per cent, with most being between 0.3 and 0.4 per cent (Table 1). The lowest Ro values were recorded for the maceral eu-ulminite A in sections 27 and 29 (0.14 and 0.15%, respec-

TABLE 1

Reflectance and maceral composition data of the samples studied

Sample No.	Distance from top of section(m)	Type	% Macerals (by volume)						
			Ro	HT	HC	HD	LIPT	IN	MM
Section 15									
2-15-85	24.0	carb. shale		2.5	–	–	–	–	97.5
1-15-85	27.0	carbominerite		35.6	15.6	2.0	0.3	1.0	45.3
4-15-85	45.0	carb. shale		13.3	0.3	–	–	–	86.3
5-15-86	47.0	coal		39.6	17.6	32.3	2.2	1.3	6.6
6-15-85	143.0	carbominerite		45.0	8.3	1.3	0.3	0.3	44.6
7-15-85	156.0	carb. shale		1.6	–	–	–	1.0	97.3
8-15-85	259.0	carbominerite		56.6	1.6	1.0	6.0	0.6	34.0
9-15-85	397.0	carb. shale		6.3	–	–	–	1.6	92.0
10-15-85	517.0	coal	0.34	69.0	21.0	–	1.0	–	9.0
15-15-85	606.0	carb. shale	0.36	4.0	–	–	–	2.6	93.3
16-15-85	615.0	coal	0.38	64.3	22.3	4.3	5.9	0.3	2.6
17-15-85	622.0	carb. shale	0.38	1.6	–	–	–	1.0	97.3
18-15-85	637.0	carb. shale	0.38	11.3	–	–	–	–	88.6
Section 20									
2-20-85	27.0	carb. shale	0.47	0.6	–	–	–	0.6	98.6
3-20-85	38.0	coal	0.37	91.3	8.6	–	–	–	–
4-20-85	41.0	carb. shale	0.34	3.0	–	–	–	–	97.0
5-20-85	82.0	carb. shale	0.27	1.3	–	–	–	0.3	98.3
7-20-85	304.0	carb. shale	0.29	3.3	–	–	–	1.6	95.0
Section 27									
6-27-85	87.0	carbominerite	0.14	58.6	–	–	–	0.3	41.0
Section 28									
-28-85	4.5	carb. shale		9.6	0.6	0.3	0.3	1.3	87.6
-28-85	23.0	coal		47.3	15.3	13.6	2.9	1.0	19.6
-28-85	42.0	carb. shale		10.6	3.0	–	–	0.6	85.6
-28-85	52.0	carb. shale		18.3	19.3	3.6	5.6	–	53.0
-28-85	61.0	coal		29.3	28.6	12.0	3.6	17.6	8.6
-28-85	96.0	carb. shale		7.3	23.6	7.3	2.6	0.3	59.6
Section 29									
3-29-85	79.0	coal	0.15	98.3	0.6	–	–	0.6	0.3
Section 30									
1-30-85	24.0	carb. shale	0.23	4.3	3.3	–	0.3	0.6	91.3
4-30-85	56.0	coal	0.23	49.0	42.0	1.3	–	–	7.6
5-30-85	93.0	carb. shale	0.29	4.0	2.3	5.3	3.6	1.0	83.0

HT, humotelinite; HC, humocollinite; HD, humodetrinite; LIPT, liptinite; IN, inertinite; MM, mineral matter.

tively), whereas the highest value was recorded in section 20 (0.47% for eu-ulminite B). Reflectance values indicate a low stage of coalification (lignite) for the fossil forest coals.

Fluorescence measurements (λmax and R/G quotient) were taken for the macerals resinite, sporinite, cutinite, suberinite, exsudatinite, and fluorescing huminite (wood) (Table 2) (Fig. 2). Fresh resinite has a λmax range between 430 and 510 nm (R/G Q = 0.1-0.36), whereas oxidized resinite has a λmax of 530 to 540 nm and R/G Q of 0.53-0.63 (Fig. 2a). Fresh sporinite has a λmax of 440 to 500 nm (R/G Q = 0.26-0.50) and oxidized sporinite has a λmax of 540 nm (R/G Q = 0.73) (Fig. 2b). Fresh cutinite has a λmax range of 440 to 500 nm (R/G Q = 0.12-0.57) and oxidized cutinite has a λmax of 580 to 610 nm (R/G Q = 0.90-1.65) (Fig. 2c). Suberinite has a λmax of 440 nm (R/G Q = 0.12) (Fig. 2d) and exsudatinite has a λmax of 440 to 460 nm (R/G Q = 0.22-0.31) (Fig. 2e). Fluorescing huminite has a λmax of 450 to 470 nm (R/G Q = 0.41-0.54) (Fig. 2f). Fluorescence results (i.e., low λmax and R/G Q values) corroborate with those obtained from reflectance, thus verifying the low rank of the fossil forest coals.

DEPOSITIONAL ENVIRONMENT

At the fossil forest site, fining-upward siltstone-coal sequences represent successive crevasse splays and more distal flood basin sediment accumulations during periods of flooding (Ricketts, *this volume*).

Peat accumulation can occur in the floodplains associated with channel systems (Jerzykiewicz and McLean, 1980). The prerequisites for accumulation of thick peat deposits are: 1) an

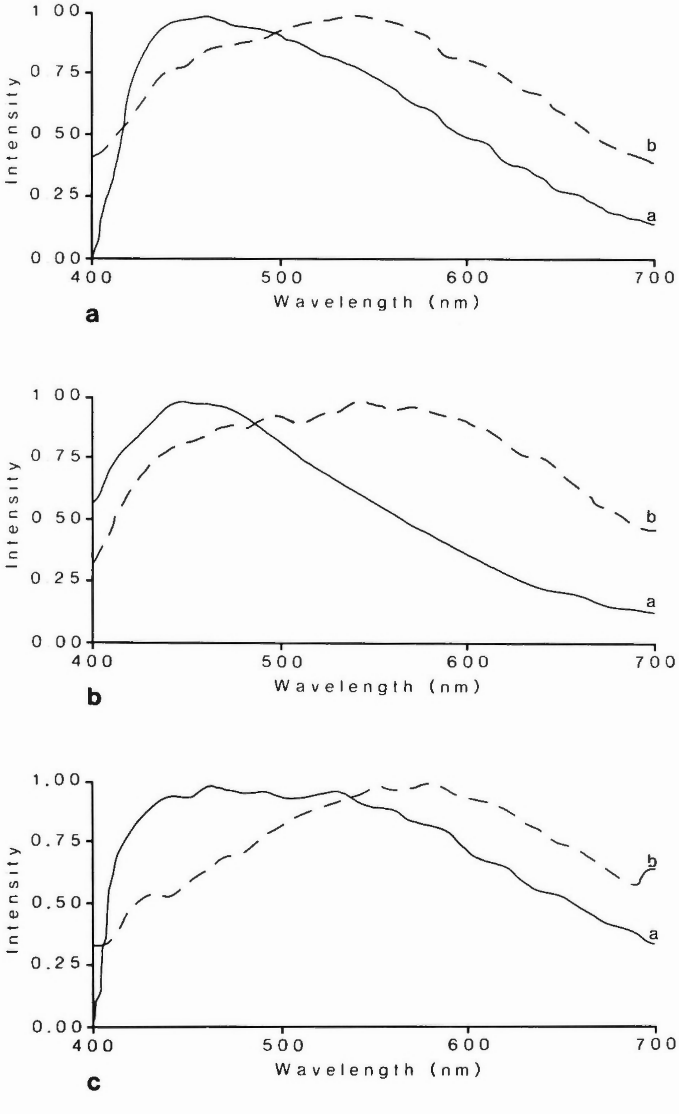

Figure 2*. Fluorescence spectra of liptinite macerals. a, resinite, fresh (—) and oxidized (– – –); b, cutinite, fresh (—) and oxidized (– – –); c, sporinite, fresh (—) and oxidized (– – –); d, suberinite; e, exsudatinite; f, fluorescing huminite.*

TABLE 2

Spectra fluorescence parameters (λmax, R/G Q) of liptinite macerals from selected samples

	λmax (nm)	R/G Q
Resinite, fresh	430-510	0.10-0.36
Resinite, oxidized	530-540	0.53-0.63
Sporinite, fresh	440-500	0.26-0.50
Sporinite, oxidized	540	0.73
Cutinite, fresh	440-500	0.12-0.33
Cutinite, oxidized	580-610	0.90-1.65
Suberinite	440	0.12
Exsudatinite	440-460	0.22-0.31
Fluorescing huminite	450-470	0.41-0.54

adequate supply of plant matter; 2) prolonged maintenance of a precise balance between the groundwater level and the peat surface; and 3) little or no influx of detrital clastic sediment (Teichmüller, 1982).

The depositional environment for the fossil forest coal is envisaged as a floodplain associated with an active channel (30-50 m minimum width) (Ricketts, *this volume*). Much of the meander plain was covered by forest type vegetation, with peat formation interrupted during major flooding events. The flood events are represented by mudstone layers among the coal seams.

The coniferous trees developed in areas remote from overbank flooding, in a hospitable environment for mixed conifer and hardwood forests and subsequent peat accumulation. In contrast, favourable conditions could not develop in the braidplain where vegetation was sparse, flooding was more frequent (2-4 m water depth) (Ricketts, *this volume*) and, as a result, woody debris (humodetrinite) and mineral matter are more common.

A balance was evidently achieved between water levels, rate of subsidence and accumulation of vegetal matter during peat deposition, so that relatively thick (<1.5 m) and clean coal beds with only minor quantities of mineral matter were formed. The excellent preservation of cell structure, distinct floor and roof boundaries, presence of seat earth, thickness and continuity of the seam (2-5 km) (Ricketts, *this volume*) all tend to indicate a high degree of autochthony (Teichmüller, 1982; Goodarzi and Gentzis, 1987). The presence of partings indicates unfavourable conditions for peat accumulation.

An attempt has been made to establish a correlation between coal facies indicators and the environment of coal formation. A slight modification of Diessel's coal facies diagram for a variety of depositional settings (Diessel, 1986) has been attempted with limited success. The model takes into consideration two parameters, the gelification index (GI) and tissue preservation index (TPI) (Fig. 3). GI and TPI are calculated from the maceral data (mmf) of the coals only as follows:

$$GI = \frac{\text{total huminite and macrinite}}{\text{semifusinite, fusinite and inertodetrinite}}$$

$$TPI = \frac{\text{humotelinite, semifusinite and fusinite}}{\text{humocollinite, humodetrinite, macrinite and inertodetrinite}}$$

Most of the samples fall within the wet forest, fen and telmatic environments. The fossil forest coals have relatively high gelification indices, an indication that they have a high content of wood-derived structured huminite (humotelinite), and a low inertinite content. When total huminite is dominated by humotelinite, this results in high tissue preservation indices (Diessel, 1986; Fig. 3). On the other hand, when unstructured huminite (humocollinite and humodetrinite) prevail, TPI values are low.

Ternary composition diagrams

A more useful alternative to Diessel's method may be the ternary diagram where mineral matter occupies one apex and groupings of the three maceral groups occupy the others. The diagram, which is a modification of one used by Goodarzi (1985) and by Goodarzi and Gentzis (1987), is shown in Figure 4. The fossil forest coals plot mainly in the region that

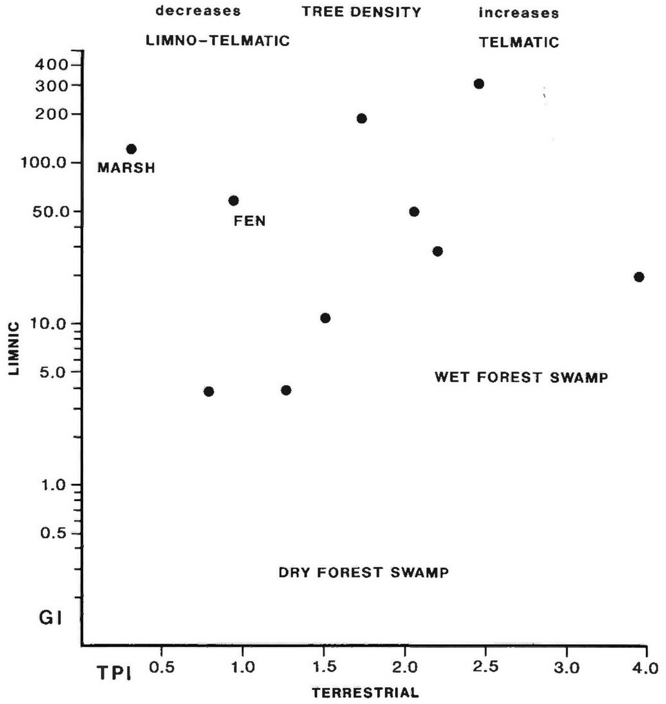

Figure 3. Gelification versus tissue preservation index diagram. (Modified from Diessel, 1986.)

indicates good preservation conditions in an environment rich in tree-like vegetation (Fig. 5). The presence of tree stumps in growth position and leaves in the mudstone and coal indicates a forested flood basin depositional environment, in agreement with the interpretation by Ricketts (*this volume*). Almost all of the fossil forest coal samples contain high amounts of huminite and, generally, low amounts of liptinite and inertinite (Fig. 6). This indicates that the peat swamp rapidly achieved conditions favourable for the preservation of organic matter, and that only limited biological degradation of the woody tissues took place. An inertinite-rich interval in section 28 represents a gradual shift toward drier conditions and a corresponding increase in inertinite at the expense of humotelinite and humocollinite.

Comparison with the Hat Creek coal (British Columbia)

The petrology of the intermontane coals of British Columbia, in particular the Hat Creek coal deposit, is very similar to that of the fossil forest coal. The coals in both areas are of similar age (Eocene), contain high huminite (humotelinite and humocollinite) and low inertinite and liptinite.

Both coals are bright to banded bright, dominated by coniferous woody material (*Metasequoia*), they are autochthonous (in situ), and were deposited in a forest-type environment.

In addition, both coals formed in a fresh water environment, as indicated by low boron contents of less than 40 ppm (Goodarzi, 1987; Goodarzi and Van der Flier-Keller, 1988; Goodarzi and Hill, 1989). The difference, however, is that the fossil forest coal seams have developed from peat accumulation in a densely vegetated meander plain adjacent to a sparsely vegetated braidplain, fed by an alluvial fan. The Hat Creek deposit was developed in an intermontane fresh water (possibly lacustrine) body, slowly subsiding due to block faulting.

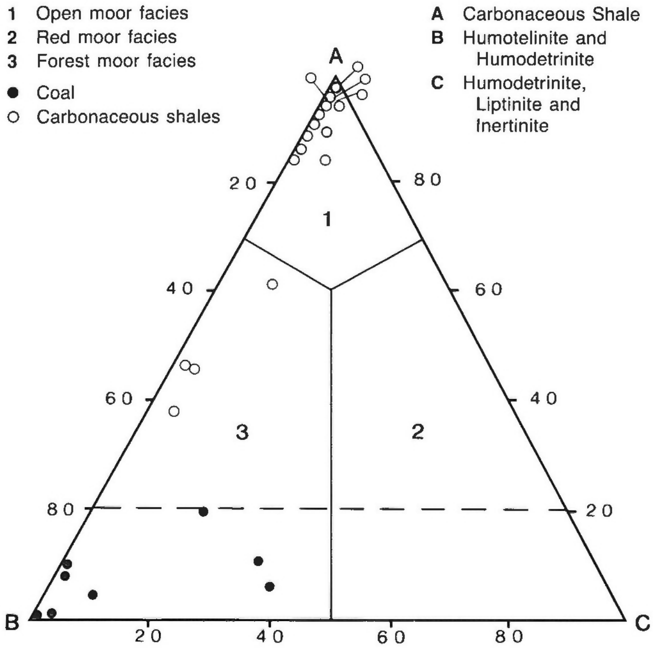

Figure 4. Ternary composition diagram showing coal facies.

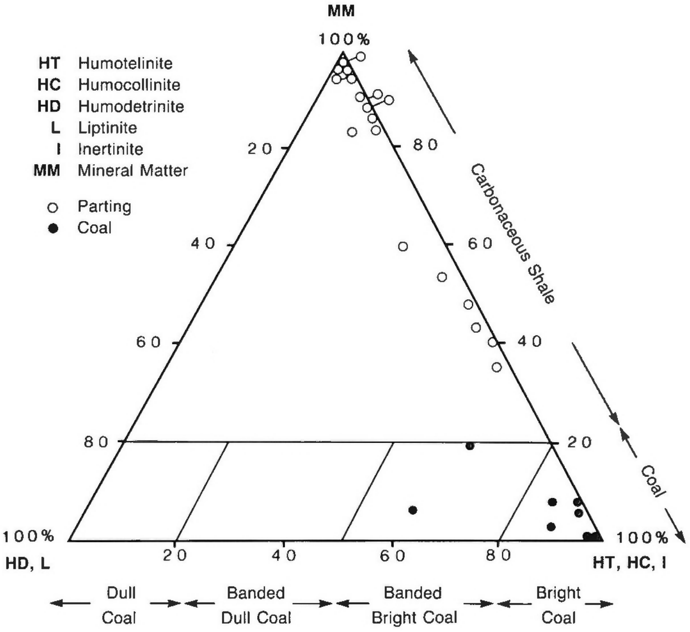

Figure 5. Ternary composition diagram of lithotypes.

The difference between the two coals is that the Hat Creek seams are much thicker (up to 20 m), compared with the thinner seams in the fossil forest (<1.5 m). This may be due to the

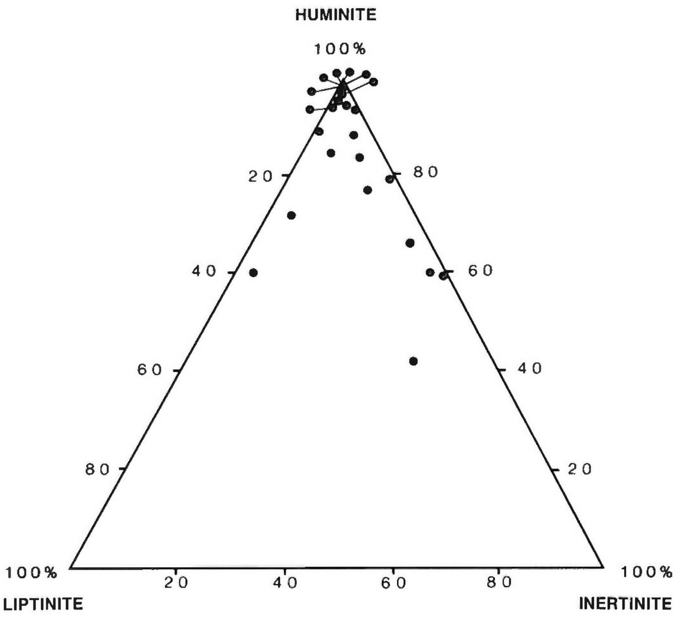

Figure 6. Ternary maceral composition diagram.

different environment of deposition, i.e., alluvial fan (fossil forest) versus lacustrine (Hat Creek). Also, the rank of the Hat Creek coal (0.35-0.45% Ro) as measured by eu-ulminite reflectance is in a narrower range than the fossil forest coal (0.15-0.47% Ro).

SUMMARY

The Middle Eocene coals from the fossil forest succession are low-rank (lignite) and dominated by huminite group macerals. Liptinite and inertinite are minor components. The coal probably formed from a forest-type vegetation developed on a meander plain. The high humotelinite content attests to autochthonous (in situ) deposition at a time when conditions were favourable for the preservation of woody tissue.

The fossil forest coal is petrographically and geochemically (from the boron content) similar to the Hat Creek deposit of British Columbia. This indicates fresh water depositional environment for the fossil forest coal.

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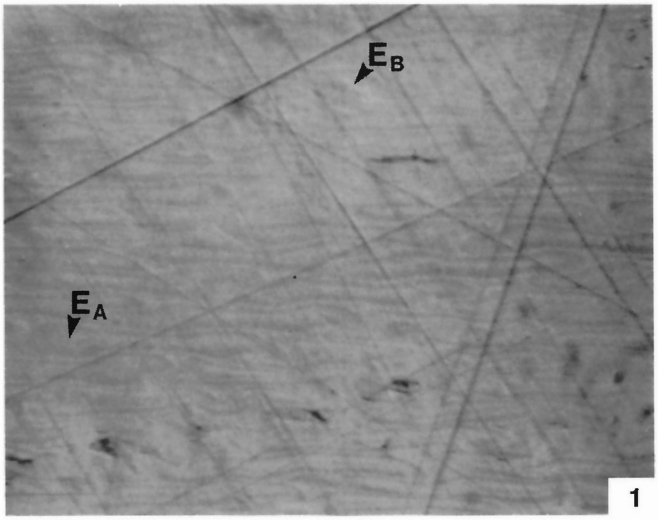
PLATE 1

All photomicrographs taken in black and white, oil immersion.
Long axis is 180 µm.

Figure 1.	Eu-ulminite A (EA) and eu-ulminite B (EB).
Figure 2.	Eu-ulminite B (EB) band in shaley matrix.
Figure 3.	Gelinite as telogelinite (TG) in association with ulminite. Note the higher reflectance of gelinite.
Figure 4.	Phlobaphinite (PH) and porigelinite (PG) association.
Figure 5.	Cutinite (CU), sporinite (Sp) and inertodetrinite (ID) in humodetrinite.
Figure 6.	Fusinite (F) and humodetrinite (HD).



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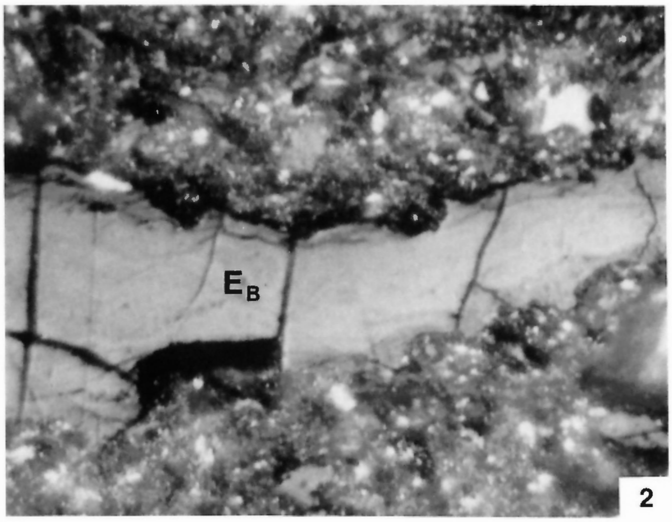

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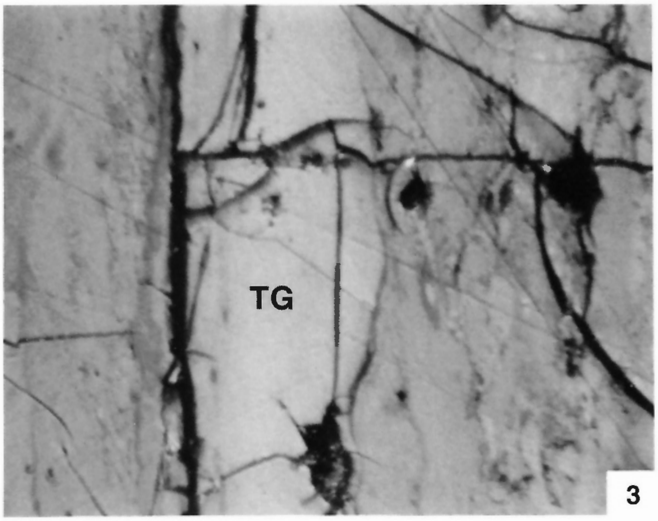

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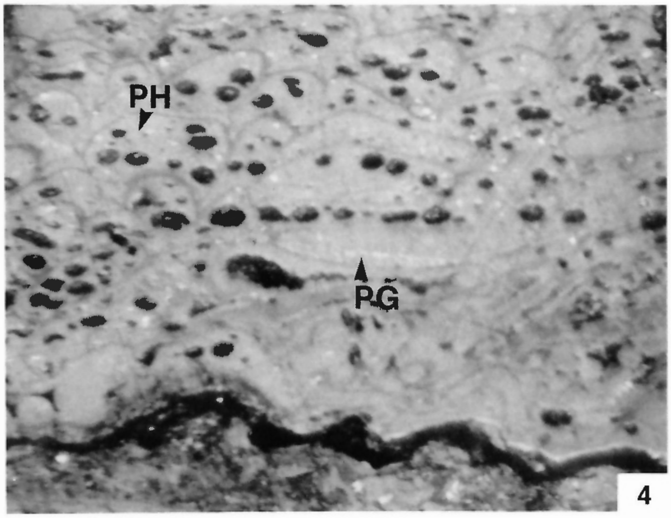

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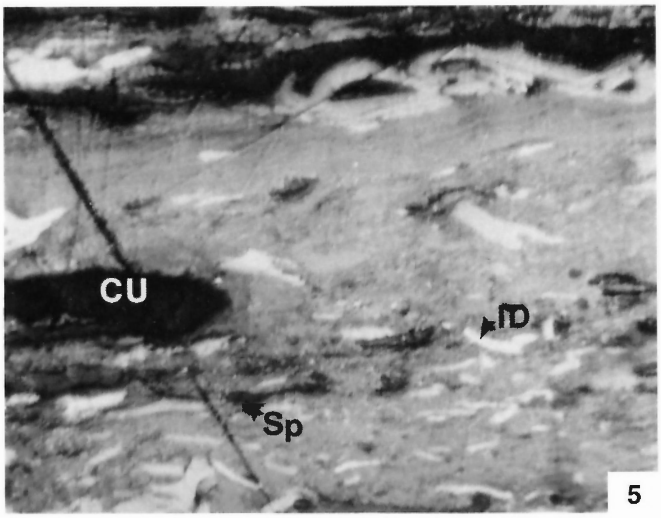

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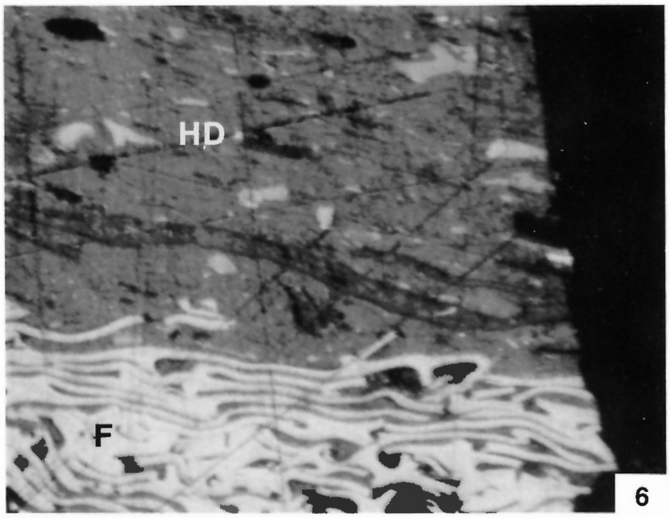

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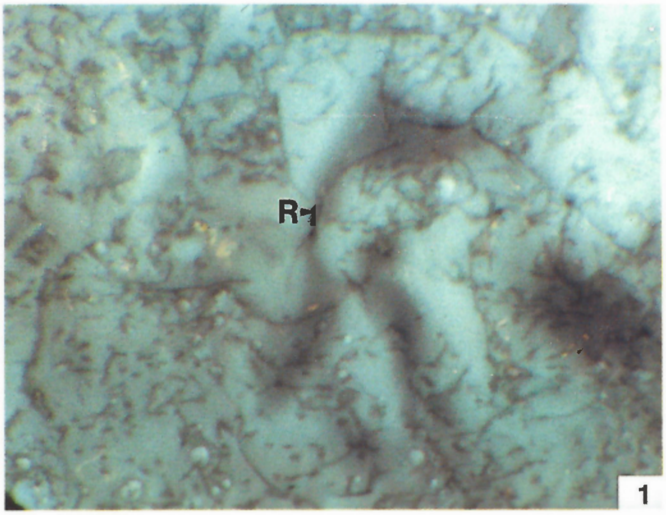
PLATE 2

All photomicrographs taken in fluorescent light, water immersion (Excitation filter 365 nm, barrier filter 420 nm). Long axis is 180 µm.

Figure 1.	Massive resinite (R) showing evidence of degration.
Figure 2.	Oval resinite (R), folded cutinite (CU) and sporinite (Sp).
Figure 3.	Thin-walled cutinite (CU) and microspores (Sp).
Figure 4.	Brightly fluorescing body (alginite?) (A) in huminite matrix.
Figure 5.	Suberinite (Su) associated with phlobaphinite (PH).
Figure 6.	Spores (Sp) in shaly matrix.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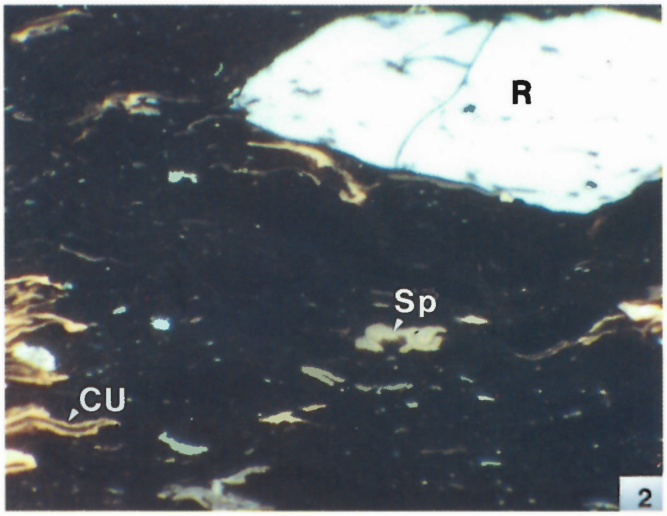
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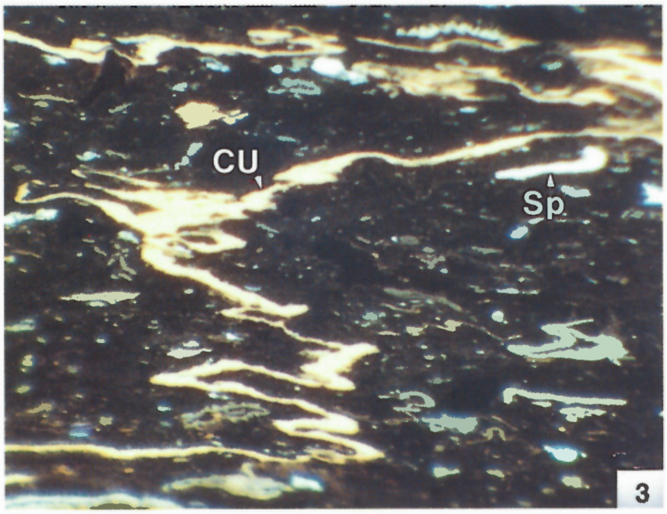
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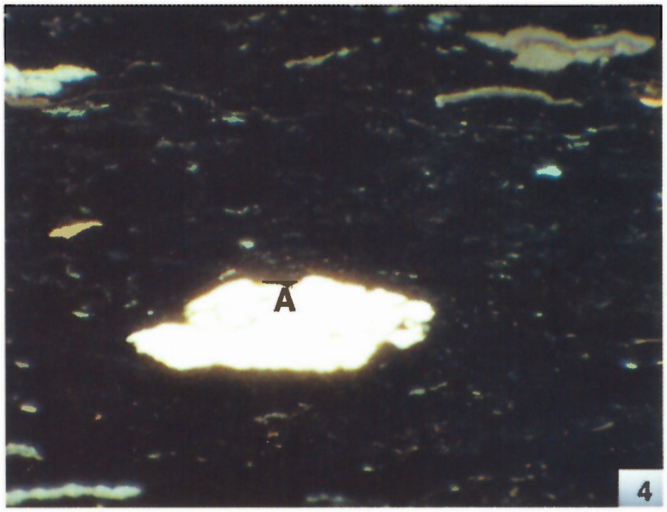
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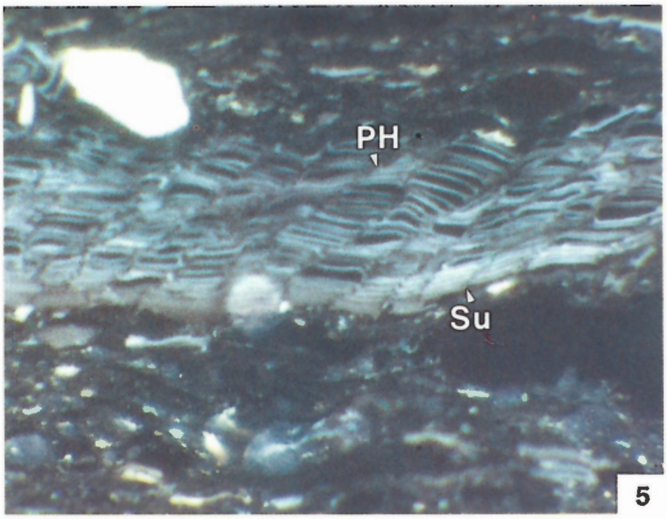
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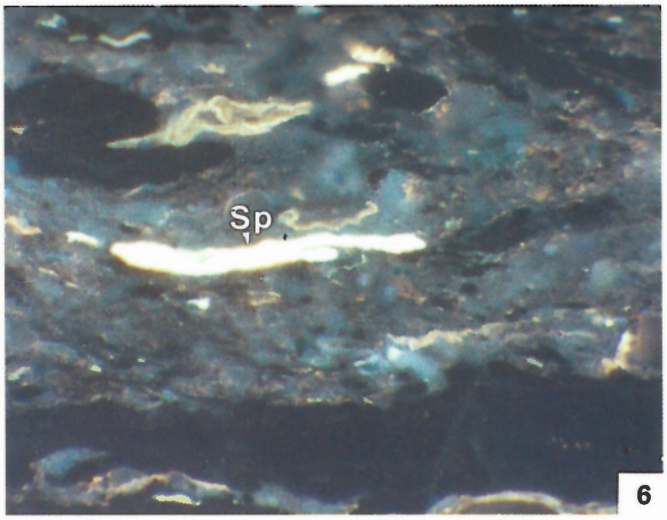
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MICROSCOPIC CHARACTERIZATIONS OF FOSSIL WOOD FROM GEODETIC HILLS, AXEL HEIBERG ISLAND

G.S. Young[1]

Young, G.S., 1991. Microscopic characterizations of fossil wood from Geodetic Hills, Axel Heiberg Island. In Tertiary Fossil Forests of the Geodetic Hills, Axel Heiberg Island, Arctic Archipelago, R.L. Christie and N.J. McMillan (eds.); Geological Survey of Canada, Bulletin 403, p. 159-170.

Abstract

Specimens of wood from two tree stumps excavated from a fossil forest site of Eocene age, near the Geodetic Hills on Axel Heiberg Island, were characterized by polarized light microscopy, phase contrast microscopy, and fluorescence microscopy. Identifications of Taxodiaceae and Pinaceae are consistent with the preserved anatomical detail. Extreme compaction and deterioration of the two samples of wood prevent possible genus identification. Growth increments are distinct, due in part to an abundance of latewood, and, therefore, are similar to those of extant genera that undergo seasonal growth between periods of dormancy. Compression has reduced the dimensions of growth increments by from 50% to over 80%, the compaction occurring mostly in the radial plane. Most of the air space in the cell cavities of the wood is lost. Much less longitudinal collapse has occurred. Overall deterioration is similar between the specimens of wood and is suggestive both of rapid burial at the time of death of the trees and of abiotic breakdown during burial, due possibly to acidic burial conditions. Some biodeterioration is evident in isolated latewood regions in the Pinaceae specimen. The presence of voids in these areas, where cellular material once occurred, indicates that at least some of this deterioration is more recent than the compaction and general breakdown of the wood substance. Possible traumatic resin canals, suggestive of frost or wind damage, also occur in the Pinaceae sample.

Résumé

Le bois de deux souches d'arbres extraites d'un site de la forêt fossile de l'Éocène, près des collines Geodetic de l'île Axel Heiberg, a été caractérisé au moyen de la microscopie en lumière polarisée, de la microscopie en contraste de phases, et de la microscopie à fluorescence. L'identification des *Taxiodiaceae* et *Pinaceae* concorde avec l'observation des détails anatomiques préservés. Le compactage et la détérioration extrêmes des deux types de bois empêchent une identification des genres possibles, mais les anneaux de croissance sont bien visibles, partiellement grâce à l'abondance de bois d'automne, donc sont semblables à ceux des genres actuels qui subissent une croissance saisonnière entre les périodes de dormance. Dans les cavités cellulaires des deux bois, la majeure partie des espaces occupés par l'air ont disparu. L'affaissement longitudinal est beaucoup moins important. Dans l'ensemble, la détérioration est la même pour les deux bois, et suggère à la fois un enfouissement rapide des arbres au moment de leur mort, et une décomposition abiotique durant l'enfouissement, dans un milieu peut-être acide. Dans les spécimens de *Pinaceae*, un certain degré de biodétérioration est évident dans des régions isolées de bois d'automne. La présence de vides dans ces zones, où se trouvait autrefois du matériel cellulaire, indique qu'au moins une partie de cette détérioration est plus récente que le compactage et la décomposition générale de la substance ligneuse. On rencontre aussi dans l'échantillon de *Pinaceae* des canaux résineux qui sont peut-être le résultat d'un traumatisme, causé par le gel ou par le vent.

INTRODUCTION

Microscopic study was undertaken on two fossil tree stumps of Eocene age from the east side of Fossil Forest Hill near the Geodetic Hills on Axel Heiberg Island. The low mineral content (Grattan, *this volume*; Obst et al., *this volume*) and the cohesive property of the woods made microtomy possible, and the production of thin unembedded sections permitted the use of three microscopic techniques commonly used in studies of archaeological wood: polarized light microscopy, phase contrast microscopy, and fluorescence microscopy using blue incident illumination (~400 nm).

The microscopy was undertaken to document the distortion and deterioration of the wood anatomy and to assist in the development of conservation methods for preserving the two excavated, water saturated stumps (Grattan, *this volume*). Conservation of waterlogged wood often entails infiltrating the microscopic structure with a polymer, e.g., polyethylene glycol. This has the effect of maintaining the water swollen

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[1] Canadian Conservation Institute, Department of Communications, 1030 Innes Road, Ottawa, Ontario K1A 0C8

dimensions of the wood and of preventing shrinking and cracking when the wood dries. Knowledge of the type of wood and its deterioration is important for designing suitable treatment (Young and Sims, 1989); microscopic study provides this information. Such studies were therefore undertaken for the following reasons: 1) to identify the fossil wood, 2) to document the fossil morphology including lateral collapse (in the radial and tangential directions) and longitudinal collapse (in the axial direction), and 3) to assess the type and extent of deterioration. Microscopy was also used in an attempt to gather information about the environmental conditions under which the trees grew and about the circumstances surrounding their burial.

EXPERIMENTAL METHODS

Wood samples

Three samples of wood, each approximately 3 cm^3, were taken from the same tree stump, designated HR4 (Heart Rot), excavated from the east side of Fossil Forest Hill. One sample was taken from each of three pieces of the trunk stored in water. The pieces were taken from above what appeared to be buttresses on the stump, from the outer portion of the trunk, which measured up to 80 cm in diameter. The samples were therefore thought to consist of growth increments containing mature wood anatomy.

A single sample of dry wood was also collected from an air dried tree stump, designated NMNS S1 (National Museum of Natural Sciences). As with the samples from HR4, this sample was taken from the outer, mature rings of the stem wood in the stump.

Preparation for microscopy

Embedding for microtomy proceeded according to standard histological practices, first with substitution of solvents from water to ethanol to n-propanol to n-butanol. Solvent was then substituted by butoxyethanol-glycol methacrylate, and infiltration proceeded for up to two weeks with constant sample agitation. Polymerization was carried out at room temperature in capped polyethylene capsules.

Microscopic sections approximately 10 μm thick were cut from the embedded wood samples with a Porter-Blum JB-4 rotary microtome. Steel blades were used, at an angle of approximately 6 degrees to the surface of the samples (Dinwoodie, 1966). Transverse and longitudinal sections were prepared. Because of the compaction of the wood structure, little infiltration of the samples with methacrylate had occurred. However, initial attempts at sectioning revealed that the unembedded wood could be cut easily. The polymerized methacrylate blocks provided the extra support needed to permit sectioning at the stated thickness, one thin enough to permit optically undistorted microscopic examination by transmitted light. Permanent microscopic slides were made by placing the sections into a drop of proprietary resinous mounting medium ($n = 1.56$) and applying a coverslip. Pressure was applied for several days to the sections. Morphological study was carried out by microscopic examination of microtomed sections cut in the transverse, radial and tangential planes of the woods.

WOOD IDENTIFICATION

Although the fossil wood had collapsed under powerful compressive forces and had undergone extensive deterioration, many of the anatomical features of the secondary xylem have remained sufficiently intact for use in identification. Table 1 lists the features observed in the fossil wood sections.

No single anatomical feature is sufficient to identify a particular wood (Basinger, 1981), but a combination of features can usually permit the identification of a wood genus and, frequently, even a species. Wood identification is complicated by two seemingly conflicting factors: on the one hand, there is some variability of anatomical detail within species, and even within individual trees, which results from the influence of the growth environment (Hejnowicz, 1973; Wilson and White, 1986), and, on the other hand, anatomical similarities resulting from heredity are found among related and some unrelated genera, especially among softwoods. Therefore, an identification based strictly on xylem morphology, especially when it is deteriorated, often results in several possibilities.

The common morphological features of wood are useful for distinguishing taxonomic classes. For the fossil wood of this study, the absence of true vessels and the presence of rays and of tracheids in parallel radial rows allowed identification of both samples as Coniferae.

HR4

Identification of HR4 is difficult, because several important features are not observable in sections cut from the compressed, distorted wood. Features that are usually associated with the tracheids, such as the organization of bordered pits on radial walls, are totally obscured in the sections. The presence or absence of unique features, such as crassulae or trabeculae, cannot be determined, and much of the minute structural detail in ray parenchyma cells is too distorted to be useful.

Axial parenchyma, an important element in these identifications, could not be confirmed. In transverse section, the presence of clear, dark brown (resinous) deposits scattered throughout the earlywood and latewood in growth rings suggests their presence (Fig. 1), but similar deposits were also found in the lumens of some latewood tracheids (Fig. 5). In tangential section, moreover, no transverse wall was observed in cells containing these deposits.

Despite the distortion, the sections contained enough recognizable structural detail to allow a partial identification of the wood. Structural features include: 1) uniseriate rays consisting of procumbent cells (Fig. 2; no biseriate rays were encountered), 2) smooth end (tangential) walls of ray parenchyma (Fig. 3), 3) thickened and unpitted or slightly pitted horizontal walls of ray parenchyma, 4) taxodioid-like pitting in single horizontal rows in cross-fields (Fig. 4), and 5) well defined growth rings containing latewood (Figs. 1 and 5; Wilson and White, 1986, p. 213). Both transverse and axial resin canals are absent from the wood. These features are consistent with an identification of Taxodiaceae.

The presence both of delineated growth rings and of thickened, mostly smooth (infrequently pitted) horizontal walls of ray parenchyma generally precludes identification of Podocarpaceae, Cycadaceae, Ginkgoaceae and Araucariaceae (Greguss, 1955) and, in combination with smooth end walls in ray cells, also excludes Cephalotaxaceae (Phillips, 1979). The lack of axial and transverse resin canals, plus the presence of taxodioid-like pitting and smooth end walls in ray cells eliminate Pinaceae. Taxodiaceae, Taxaceae and Cupressaceae remain, and the alignment of the large, taxodioid-like cross-field pitting in single horizontal rows and the distinct growth rings (distinct because they contain conspicuous latewood) suggest association with Taxodiaceae.

Within this family there are eight extant genera: *Athrotaxis* spp., *Cunninghamia* spp., *Cryptomeria* spp., *Glyptostrobus* spp., *Metasequoia* spp., *Sequoia* spp., *Taiwania* spp., and *Taxodium* spp. Selecting an anatomically homologous genus from among these would require a view of undistorted morphological detail of the fossil wood, particularly in ray parenchyma cells. Because this is not available in the microtomed sections, it is impossible to identify the specimens more precisely.

NMNS S1

Many of the features noted for HR4 are also distorted beyond recognition in NMNS S1, but, in addition, no arrangement of cross-field pitting is observable. The pits, partly visible in transverse section, are small and circular, but, because they could not be seen intact in the radial plane of the wood, no specific morphology, number or arrangement of the pits in cross-fields could be determined.

Apart from this, the two samples of wood are easily separated, each into a different taxonomic family. This is indicated

TABLE 1

Anatomical features of fossil woods

Anatomical feature	HR4	Dry wood
Earlywood/latewood ratios	2.4/1 – 1/1.25	5.5/1 – 7.5/1
Wood parenchyma	?	Yes
End walls of wood parenchyma	?	Nodular
Resin deposits (longitudinal)	Distributed diffusely throughout growth increments	Few
Transition from earlywood to latewood	Appears gradual	Appears gradual
Traumatic resin canals	No	?
Normal resin canals	No	No
Fusiform rays	No	No
Latewood tangential wall pitting	Yes	?
Ray bundle height	1-10+ cells	5-20+ cells
Ray cell type	Procumbent	Procumbent
Ray bundle type	Uniseriate	Uniseriate
Ray cell shape (tangential view)	Isodiametric in earlywood and isodiametric to oval in latewood	Mostly isodiametric
Ray cell end walls	Smooth, mostly vertical	Nodular
Ray cell end walls indentured	?	?
Ray cell horizontal walls	Thickened, some pitting	Thickened, moderately pitted
Ray cell intercellular spaces	Yes	No
Ray cell cross-field pitting	Taxodioid, 1 to 3 in one horizontal row	?

by the presence in NMNS S1 of true secondary cell wall pitting in the end walls (i.e., nodular end walls) of the ray cells (Fig. 6). (HR4 has smooth, unpitted end walls in rays.) For NMNS S1, this pitting, plus the pitting visible in the transverse walls of axial parenchyma and the distinct bands of latewood (Fig. 7), exclude Taxodiaceae and others as likely identifications, except Pinaceae.

Additional support for this identification comes, again, from the ray end wall pitting. *Juniperus* spp. and *Libocedrus* spp. of Cupressaceae and genera, e.g., *Abies*, of Pinaceae contain nodular end walls (i.e., true pitting of the secondary wall) in both axial and ray parenchyma (Barefoot and Hankins, 1982), just as is noted in NMNS S1. The pitting of the ray end walls in the fossil wood does not match the often scalariform-like pitting seen in *Juniperus* spp. and *Libocedrus* spp. (as seen in tangential longitudinal section) but does match the foraminate (sieve-like) end wall pitting in Pinaceae genera.

NMNS S1 contains evidence of axial resin canals (Fig. 7). The arrangement of these "canals" in a tangential row at the junction of two growth increments and the absence of corresponding fusiform rays (transverse resin canals within ray bundles) suggest traumatic origins (their development induced by injury). The extant genera, *Pinus* spp., *Picea* spp., *Larix* spp., and *Pseudotsuga* spp., are therefore eliminated because of the absence of normal axial canals and fusiform rays. The extant genera, *Abies* spp., *Cedrus* spp., *Keteleeria* spp., *Pseudolarix* spp., and *Tsuga* spp., remain (Phillips, 1979), and the presence of traumatic resin canals is consistent with these. Such canals develop in response to injury from excessive wind stress and wounding, and from frost cracking occurring early in the growing season (Panshin and de Zeeuw, 1970).

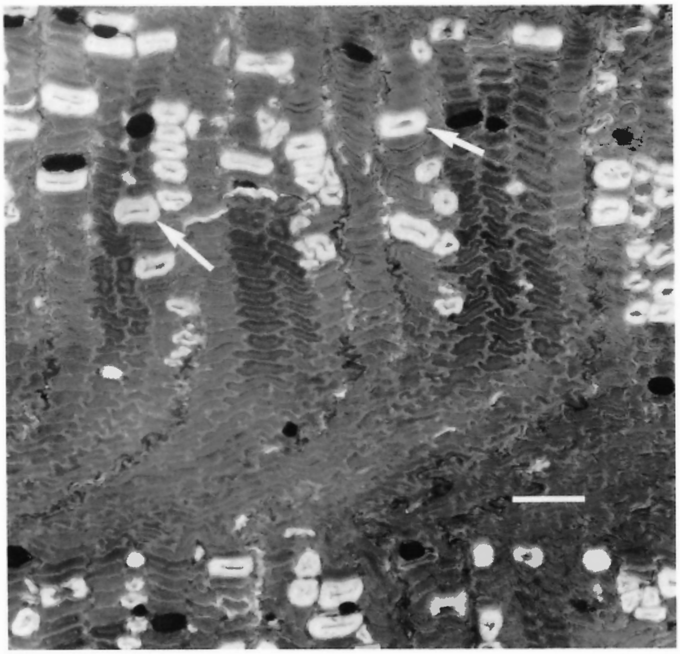

Figure 1. Transverse, brightfield. HR4 showing latewood and earlywood with lateral compaction. Many latewood cells show little or no deterioration (arrows); the majority show deterioration similar to earlywood cells, including reductions in size, darkening of colour and distortion of shape. Secondary cell walls are visible in both latewood and earlywood. (Scale bar = 75µm.)

GROWTH RING MORPHOLOGY

The extreme radial compaction of the wood samples is illustrated in the transverse sections of Figures 1, 5 and 7. Much less compaction of the tangential cell walls occurs. Collapse of earlywood is characterized by lateral sheer so that rays traverse earlywood zones in roughly straight lines, but in a direction oblique to the original tangential and radial planes. In longitudinal microscopic sections of the wood, this results in the presentation of both radial (latewood) and tangential (earlywood) cellular planes in the same section. In isolated cases, tangential compaction in the sample blocks occurs as a result of the folding of growth increments, but, again, with minimal cellular distortion in this direction (Fig. 5).

Figure 8 illustrates two portions of HR4 latewood in successive growth increments between which resides a complete,

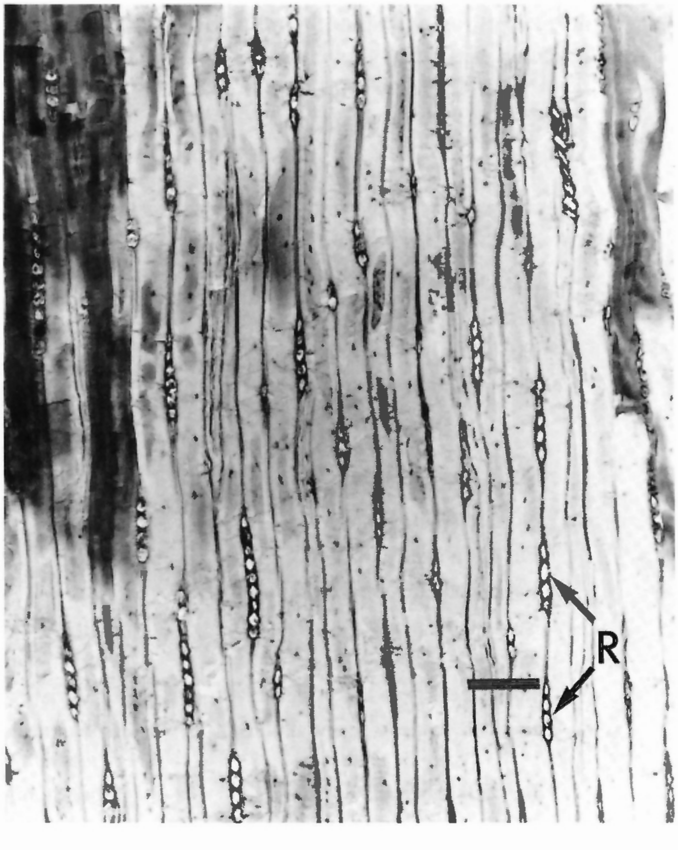

Figure 2. Tangential, brightfield. HR4 showing uniseriate rays (R). Latewood pitting on tangential walls of tracheids is visible as dark spots. (Scale bar = 115 µm.)

though collapsed, portion of earlywood. Acknowledging that cellular division occurs radially and that growth increment boundaries pass virtually perpendicular to this, the angles to which the radial orientation of the wood have been laterally distorted measure 17 degrees for the lower portion of latewood in the figure and 75 degrees for the middle section containing earlywood. Compression has reduced the original radial dimensions of the earlywood by about 75%. Combined with latewood compaction, the total for these growth increments is as much as 82%. The greater compaction of the earlywood occurs because of the greater amount of free space in the cell lumina, compared to that in the latewood, which is lost during compaction in burial. It also results from the thinner cell walls, which offer less resistance to the compressive forces in burial.

Examination of several cross-sections of the three HR4 samples reveals that earlywood and latewood compaction are variable. Different amounts of compaction can therefore be expected in other parts of the trunk wood, and measurements of 200 per cent were made from other cross-sections. Earlywood compaction for the NMNS S1 sample was also measured as high as 75 to 80%.

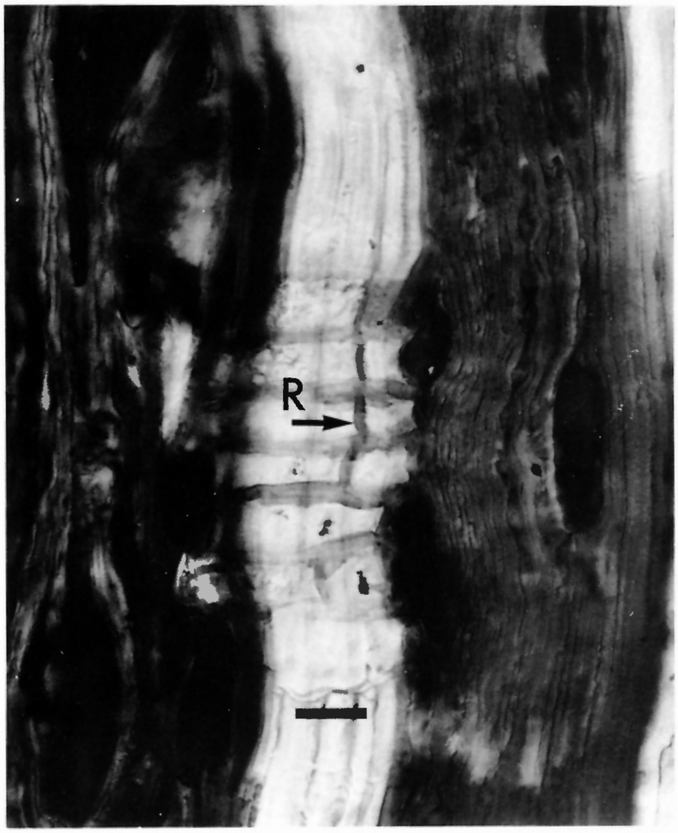

Figure 3. Radial latewood and tangential earlywood, brightfield. HR4 showing procumbent ray (R) parenchyma cells (stacked five or six cells high), containing smooth tangential walls (arrow) and thickened unpitted horizontal walls. (Scale bar = 30 μm.)

Measurements along the length of ray bundles in transverse section were used to estimate earlywood-latewood width ratios. The proportions range from 4:1 to 2:1 in the sections of the mature trunk wood of HR4, and 5:1 in the mature trunk wood of NMNS S1.

Trunk wood from large stumps usually has narrow rings, and a high proportion of each consists of latewood (Creber and Chaloner, 1984). Lower proportions of latewood occur in upper stem wood, and still lower proportions occur in branch wood and in growth rings of immature cellular development occurring near the pith. Latewood development in tree trunks is, in part, a consequence of seasonal fluctuations of the growth environment. It is also strongly influenced, however, by hormonal changes during the growing season and the increased availability of photosynthate due to reduced demands by the leaves and terminal branches after the maturation of the leaf canopy. Therefore, whereas the thickness of the latewood in a growth increment is partly influenced by environment, the presence of latewood is more a result of

Figure 4. Radial, brightfield. HR4 showing taxodioid-like cross-field pits (P) in single horizontal rows. (Scale bar = 30 μm.)

inherited physiological mechanisms (Panshin and de Zeeuw, 1970, p. 48-50; Wilson and White, 1986, p. 213).

Across a growth ring, the tensile strength increases to a maximum in the latewood (Panshin and de Zeeuw, 1970). This added strength may permit increased resistance to wind (bending) stresses.

The earlywood-latewood ratios compare well with those of many modern, temperate Pinaceae and Taxodiaceae genera. Without, however, the benefit of a broad sampling, it can at least be said that the ratios reflect adequate conditions for growth of the two trees in an environment to which they were suitably adapted to survive. Water was adequate during the entire growing season, as indicated by a lack of "false rings" and by what appears to be continuous, normal growth activity, including the development of a substantial proportion of latewood in the growth increments. The distinct boundaries between growth increments reflect seasonal growth and periods of dormancy.

Carbon dioxide and growth

Assuming that water and ground nutrients are abundant and have no limiting influence on growth, the rate of photosynthate production is mediated by temperature and light levels, but also by the availability of carbon dioxide. Within limits, increased levels of CO_2 can significantly increase the rate of growth. Kramer and Sionit (1987) reported 60 per cent more total dry weight in *Pinus taeda* (loblolly pine) seedlings grown in the presence of 0.05 per cent CO_2 compared with 0.035 per cent.

Carbon dioxide concentration and temperature are mutually compensating factors (Leverenz and Lev, 1987) and, thus, higher levels of CO_2 compensate for suboptimal temperatures. Heightened atmospheric CO_2 levels would have affected the growth of the fossil trees and may have contributed to the ability of the trees to grow in the High Arctic.

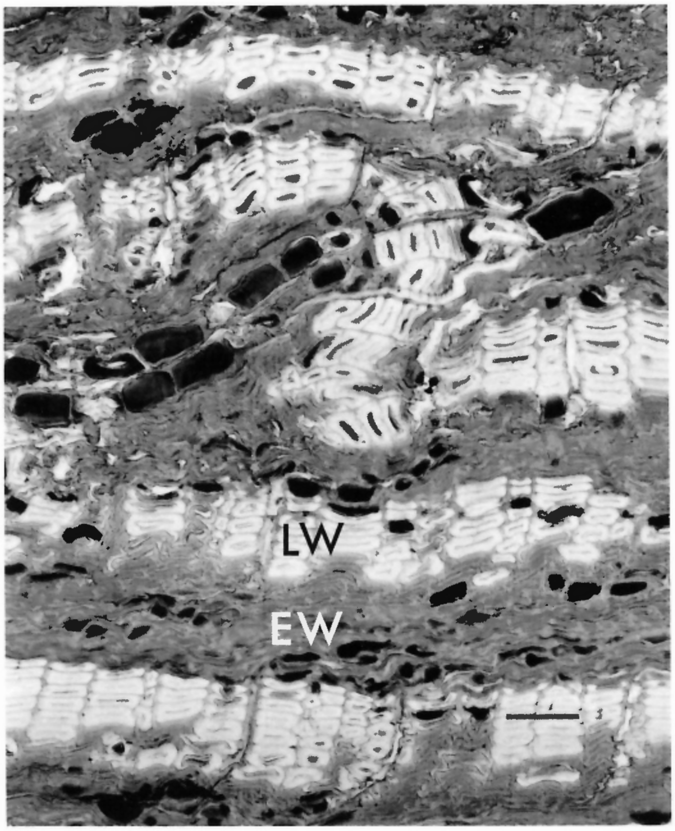

Figure 5*. Transverse, brightfield. HR4 showing relatively undeteriorated latewood (LW) and deteriorated, compressed earlywood (EW). Tangential compression occurred irregularly, involving more the distortion of the growth ring and less the compression of individual latewood cells. (Scale bar = 60 µm.)*

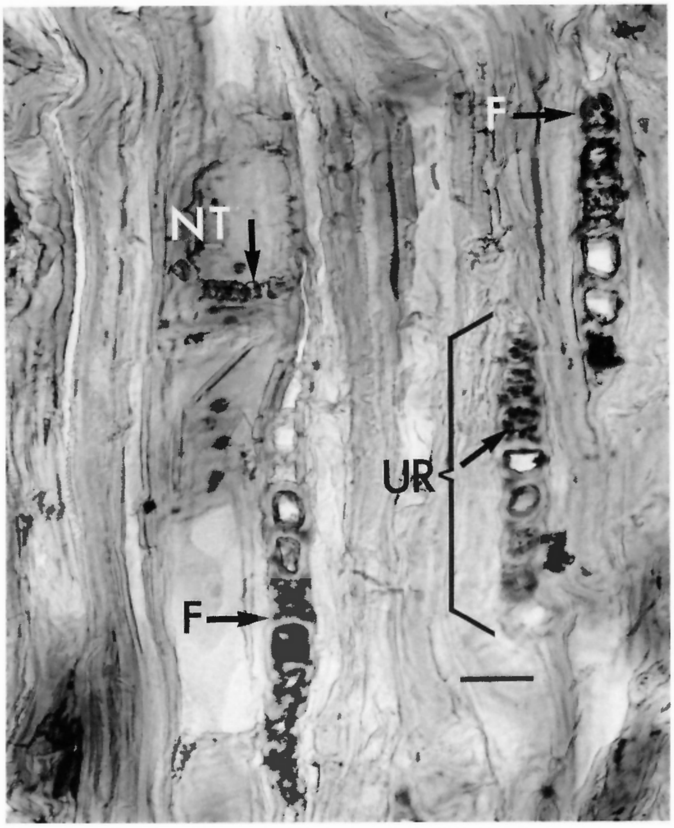

Figure 6*. Tangential, brightfield. NMNS S1 showing: 1) axial parenchyma with nodular transverse walls (NT) and with pitting leading to tracheids, 2) uniseriate rays (UR) with pitted, nodular (foraminate) tangential (end) walls (F), and 3) thickened horizontal walls of the ray cells. (Scale bar = 45 µm.)*

LONGITUDINAL COMPACTION

Less dimensional change was observed in the direction parallel to the length of the tracheids (i.e., longitudinally, along the length of tree trunks and branches). In experiments of longitudinal compression of green and seasoned wood (Keith and Côté, 1968; Keith, 1971; Dinwoodie, 1974, 1978), microscopic compressive failure took place before the wood underwent readily visible distortion. Microscopic collapse takes the form of dislocations (slip planes) in the walls as viewed in longitudinal section. Such dislocations in the fossil wood appear in crossed polarized light as dark cross-markings, shaped as X's, V's and I's, on double walls (i.e., two juxtaposed cell walls with the middle lamella sandwiched between; Fig. 9). Normally, the parallel arrangement of cellulosic microfibrils within the S2 secondary cell wall layer, and the orientation of this layer in relation to the crossed polarizing and analyzing filters in the microscope, determine the brightness of cell walls in longitudinal sections. The contrast that occurs at the dislocations results from the reorientation and ultimately the breakdown of the parallel microfibrils at these disrupted locations (Keith, 1971).

With high moisture content, as found within the buried fossil woods (Grattan, *this volume*), wood can undergo plastic-like deformation even at a compressive load only 60 per cent of that required for sudden gross failure (Keith, 1971). Dinwoodie (1968) proposed a simple, arbitrary scheme for classifying different degrees of microscopic failure. Each of four classes depends on the number of slip planes that occur and how they are arranged in proximity to each other. All of the classes are observable in the fossil wood, including the most complex (designated C_3), which comprises more than five slip planes closely situated along a single tracheid, with the same occurring in neighbouring cells.

Figure 9 illustrates C_3 slip planes. In crossed polarized light, several years of latewood appear bright in radial section. In between, regions of earlywood (seen in the tangential plane because of compaction in a direction oblique to the radial and tangential planes) show little or no birefringence. Slip planes and C_3 distortions occur throughout the latewood regions (arrows).

Vertical compressive forces acting upon the fossil woods in burial apparently occurred in the presence of horizontal forces that gave enough support to prevent total longitudinal collapse of the wood, even in the presence of numerous major slip fractures.

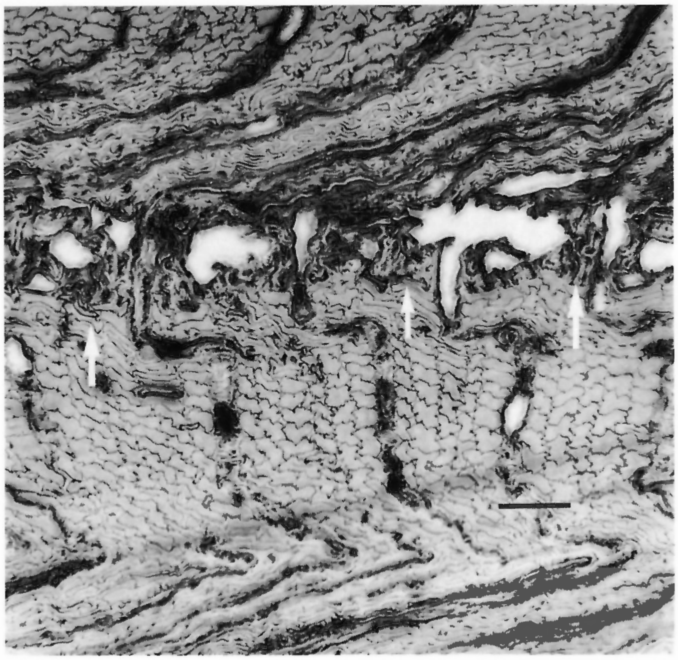

***Figure** 7. Transverse, phase-contrast. NMNS S1 showing extensive disruption of wood structure reminiscent of traumatic resin ducts in earlywood near a growth ring boundary (arrows), which traverses the photomicrograph. Remnant cell structure suggests distinct latewood zones. (Scale bar = 160 μm.)*

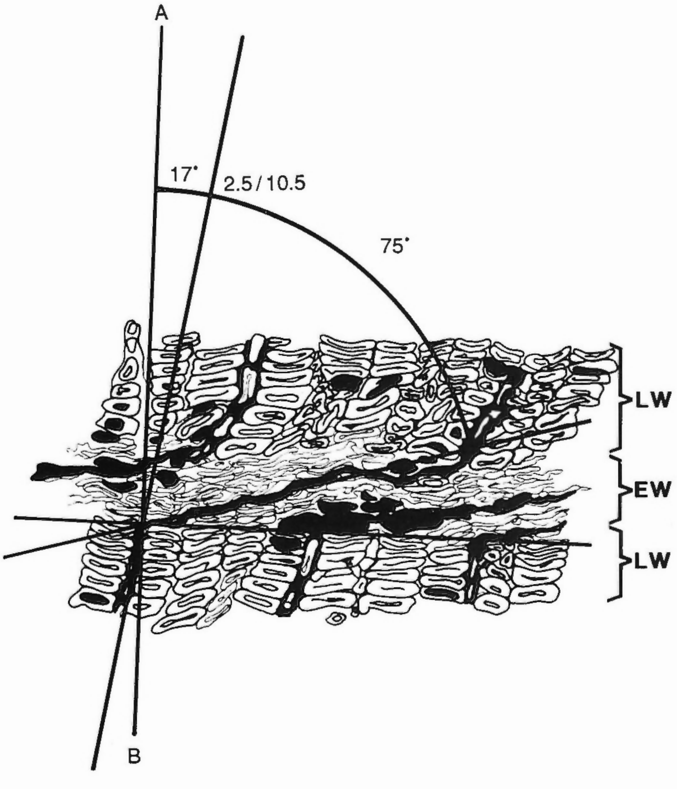

***Figure 8.** HR4. Lateral sheer 17 degrees for lower latewood and 75 degrees for earlywood, assuming AB to be original radial direction. Lateral compaction approximately 77% for earlywood (EW) and approximately 25% for the remaining latewood (LW).*

DETERIORATION

Microscopy is useful for documenting the deterioration of the ordered molecular structure and ultrastructure of wood anatomy, and it is especially useful for differentiating the various forms of microbiological attack. The manifestations of soft rot can usually be distinguished from those of brown rot. These are distinguishable from the appearance of white rot, and the manifestations of bacterial attack are often separable from these forms of fungal decay.

A morphology akin to biodeterioration is recognizable in several isolated latewood regions in transverse sections of NMNS S1 (Figs. 10, 11). The latewood tracheids show extensive breakup of the thick secondary cell wall and less of the compound middle lamella. Decay is variable among tracheids and quite localized in more intact cell walls, a feature reminiscent of the early stages of microbial attack. Whether the decay resulted from bacteria (Daniel and Nilsson, 1987; Singh et al., 1987) or fungi is, unfortunately, not discernible in this case. No similar evidence of biodeterioration was observed in HR4. Chemical analysis by Grattan (*this volume*) and by Obst et al. (*this volume*) revealed extensive loss of wood substance, particularly of the cellulosic and other carbohydrate components of the wood. Brown rot, by the Basidiomycetes, can cause such losses, but no evidence was seen of these decay organisms.

In fact, except for the isolated occurrences of biodeterioration, the microscopic evidence typified extensive abiotic deterioration. This is discussed with reference to latewood, the cell structure of which is easily recognized in the photomicrographs, but the conclusions are equally valid for the greatly distorted earlywood. As seen in Figure 1, latewood tracheids are of two types. Some tracheids appear more or less undegraded with little alteration of original shape. Most tracheids are extensively altered, as indicated by a reduction in size, darkening of colour and distortion resulting from compression.

Polarized light microscopy revealed fundamental structural differences between the two types of tracheid. The undegraded cells, designated for this study as "type one", retain their optical birefringence. The degraded cells, designated here as "type two", produce a weak birefringence limited to small portions of the cell walls.

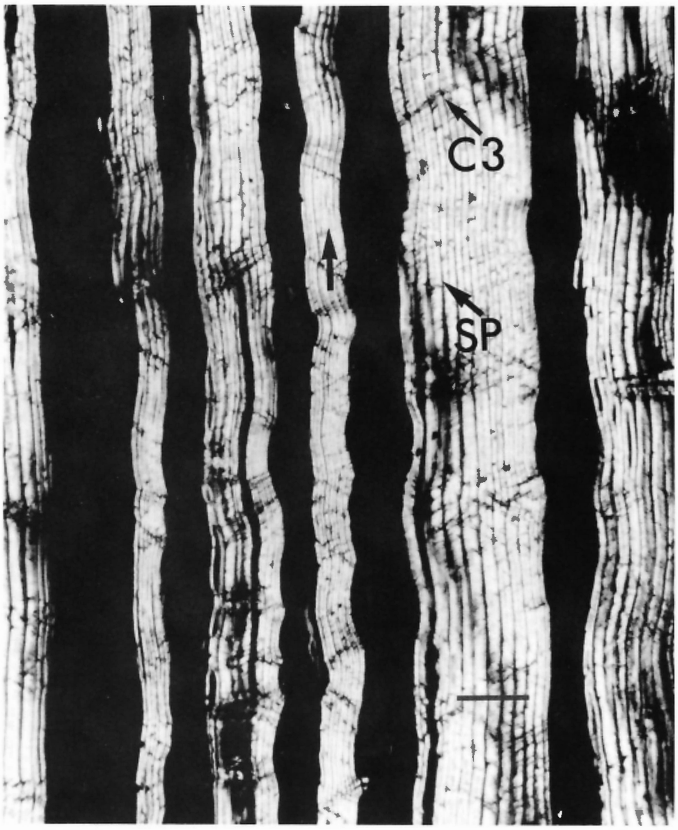

***Figure 9**. Radial latewood, polarized light, crossed polarizers. HR4 showing slip planes (SP) on individual tracheids and C_3-type fractures (C_3) across neighbouring tracheids. (Scale bar = 115 μm.)*

***Figure 10**. Transverse, brightfield. NMNS S1 latewood "type one" tracheids showing pitting (P) and the resulting breakup of the secondary cell wall indicative of biodeterioration. (Scale bar = 30 μm.)*

Permineralization

Preliminary tests were performed to investigate whether some component of the form birefringence of the two types of latewood tracheid results from cellulose microfibrils or from the very beginning of permineralization. Polarized light microscopy was carried out on HR4 cross-sections subjected to oxygen plasma microincineration to determine if any birefringent structure (n = 1.54 for denticulate silicification) remains in the delicate ash, composed of the transfixed inorganic component of the woody cells (Frey-Wyssling, 1976). No birefringence arose from the ashed, remnant cell structure, and this suggests that no permineralization detectable by microscopy resides in place of microfibrillar structure. Further study is needed to confirm this and to investigate the possible occurrence of other noncellulosic natural and foreign components in ordered arrays in the fossil wood ultrastructure. For the present study, it is assumed that the birefringence exhibited by the fossil wood is the result of ordered microfibrillar structure.

Abiotic deterioration

The intact "type one" cells show the expected variation in birefringence among the cell wall layers (Fig. 12; Wilson and White, 1986). The thick S2 layers have lower birefringence in transverse section compared to that at the outer circumference of cell walls at the location of the primary wall and the S1 layer of the secondary wall. (Variations in brightness are due to differences in the angle of packing of the helically arranged cellulosic microfibrils in the various layers of the wood cell wall.) Except for some degree of radial compaction, "type one" latewood cells in the fossil woods are indistinguishable in appearance from latewood cells in new wood.

The "type two" deteriorated latewood cells show a very different optical character. Much of the secondary walls appears without birefringence, indicating rather complete breakdown and loss of microfibrils and crystalline cellulose. However, the periphery of these cells still exhibits a faint birefringence corresponding in location to the primary wall and S1 (Fig. 12). Lower birefringence is also displayed by many of these tracheids at the boundary of the cell wall and the lumen, the location of an S3 layer in many species of wood.

"Type two" latewood, though quite degraded, retains much of its original organization. Examination by phase contrast microscopy reveals that kindred tracheids (those aligned in radial rows) retain a uniformity of size and that secondary walls frequently retain a uniformity of thickness around lumina (Fig. 1). Little checking, loss, or disproportionate thinning of

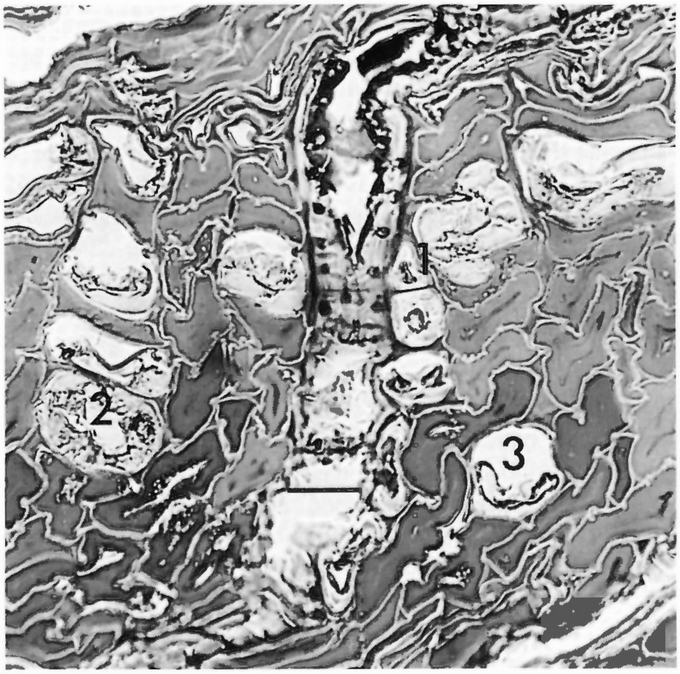

Figure 11. Transverse, brightfield. NMNS S1 latewood "type one" tracheids at various stages of biodeterioration: 1) limited attack of secondary wall, 2) attack encompassing all of tracheid, and 3) attack near completion with loss of most of the cell wall substance. (Scale bar = 30 μm.)

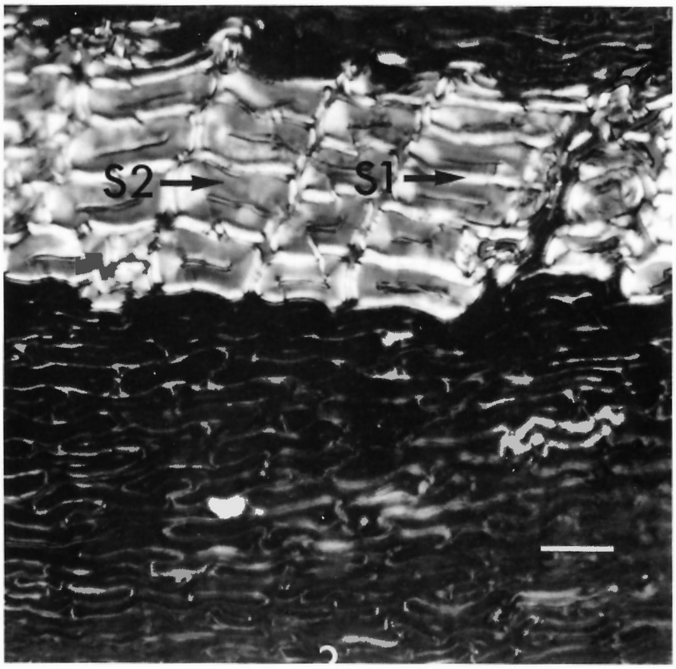

Figure 12. Transverse, polarized light, crossed polarizers. HR4 showing normal birefringence of latewood tracheids, with the S1-primary cell-wall region appearing the brightest and the S2 layer dimmer. "Type two" latewood tracheids showed a drastic reduction in birefringence, but with remnant anisotropy in the region of the primary wall and S1. Very weak birefringence occurred also at the cell wall-lumen boundary. (Scale bar = 30 μm.)

secondary walls occurs (Wilcox, 1968). Polarized light microscopy reveals that the residual birefringence of the cell wall layers also survives rather intact, showing no pitting and little loss of continuity where it remains in the cell wall (Fig. 12). Fluorescence microscopy reveals that the autofluorescent component of the lignin is also in a natural cellular organization. Compound middle lamellae have a higher concentration of lignin and therefore appear brighter than the secondary cell walls, which have a lower lignin concentration. As revealed with phase contrast microscopy, no breakup of "type two" cell wall structure is evident by fluorescence microscopy, even though the cells are reduced in size and distorted due to compressive forces.

Most forms of microbial attack would not have left such a well ordered, near normal arrangement of residual cell structure. Figure 11 shows "type one" latewood cells at various typical morphological stages of biodeterioration. Initial attack was localized, but as deterioration proceeded, the entire cell was consumed, and a small, compact residue (unrecognizable as cell wall material) was left behind (Courtois, 1966; Boutelje and Bravery, 1968; Greaves, 1969). The presence of free space in place of these cells suggests that the biodeterioration was a more recent occurrence than the abiotic deterioration and compaction prevalent throughout the bulk of the wood.

No evidence of fungal attack was seen; no hyphal bore holes were observed. Even bacterial attack is not evident, because contact-erosion troughs are absent. Localized erosion often occurs from lumina through to the S3, S2, S1 and primary wall layers (Greaves and Foster, 1970) or as tunnelling within cell walls (Daniel and Nilsson, 1987; Singh et al., 1987). Under long term exposure to bacterial attack, total loss of the secondary wall can be expected (Greaves, 1969), but, as mentioned, the fossil wood samples contained little evidence of such losses.

Some bacterial enzymes can diffuse through the secondary cell wall to cause loss of birefringence without appreciable thinning (Greaves and Levy, 1965; Boutelje and Bravery, 1968; Greaves, 1969). This can result in cells appearing similar to the degraded "type two" cells in the fossil wood during part of the deterioration process. This appearance is, however, not a common morphological characteristic of bacterial degradation, and is just one of many expected from bacterial attack.

Except for the latewood portions of NMNS S1, which showed physical disruption indicative of microbial attack (probably bacterial), the deteriorated wood cells appear to have undergone only compression and some delicate process that dissolves nonligneous components. Hydrolysis in an acidic burial environment may have been responsible. For the bulk of the sectioned wood samples, little other deterioration seems to have occurred.

Additional interpretation surrounding burial of the woods is possible. Burial of the trees must have occurred rapidly at least to the present height of the fossil stumps just prior to, or soon after, death. Because microbial deterioration of wood occurs best at ground level, partly buried, dead standing trees would have rotted at the new, higher ground level, leaving portions of the buried trunk free from attack. With the remaining tree falling to the ground because of rot, biodeterioration would continue rapidly to consume the tree unless it was also suddenly buried. Burial conditions restricted biodeterioration, probably in being anaerobic.

CONCLUSIONS

Easily cut samples from two fossil tree stumps permitted the study of wood morphology in transverse, radial and tangential microscopic sections. Tree stumps HR4 and NMNS S1 were tentatively identified as Taxodiaceae and Pinaceae, respectively. Additional microscopy of selected wood samples may provide the extra information needed to complete the identifications.

Both types of wood had undergone substantial compaction, most of which had occurred perpendicular to the axial direction of the trunk. In the presence of horizontal compressive forces in burial, longitudinal compaction of the wood has been restricted, although vertical compression is shown by an abundance of microscopic slip planes in the cell wall structure. Lumens are collapsed, and negligible free space remains. Estimates of the original thickness of the growth rings are similar to those of modern Pinaceae and Taxodiaceae genera that grow in temperate climates. The abundant latewood indicates that the two fossil trees had grown in conditions where water was at least adequate for continuous, normal, seasonal growth activity. Distinct growth ring boundaries indicate a seasonal period of dormancy. Traumatic resin canals in NMNS S1 reflect sudden stress from external forces, wind, wounding or frost.

Deterioration appears to have resulted mostly from abiotic agents, although local areas of latewood in NMNS S1 show morphology typical of microbial attack, as well as evidence that this attack occurred more recently than the overall breakdown and compaction of the wood. The general lack of biodeterioration suggests rapid burial of the stumps in anaerobic conditions. Additional study is needed to confirm the local decay by microbes and to verify that microbial attack mostly involved the comparatively intact tracheids, not the smaller, darker, distorted wood cells that make up most of the wood.

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PALEOSOLS OF THE FOSSIL FOREST AREA, AXEL HEIBERG ISLAND[1]

C. Tarnocai[2] and C.A.S. Smith[3]

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Abstract

The Eureka Sound paleosols occur in the Eocene Buchanan Lake Formation in the fossil forest area of Axel Heiberg Island in the Canadian Arctic. The study included the examination of a stacked sequence of these paleosols in a 22 m section containing at least 15 paleosols. The Eureka Sound paleosols meet the criteria for present-day Podzolic, Gleysolic and Organic soils and are found beneath contemporary Cryosolic soils. The paleosols, which all supported forest vegetation, developed on fluvial and organic material and are stacked on top of each other as a result of burial by fluvial deposition. After burial, they were affected by mineralizing groundwater, by compaction resulting from loading by sediments and glacial ice, and by faulting due to loading and tectonic movement. The data obtained during this study are useful for reconstructing the climatic, geological and pedological conditions that existed during the Eocene.

Résumé

Les paléosols d'Eureka Sound apparaissent dans la formation de Buchanan Lake, d'âge Éocène, dans la région de la forêt fossile de l'île Axel Heiberg (Archipel arctique canadien). Cette étude comprenait l'examen d'une séquence verticale de ces paléosols dans une coupe stratigraphique de 22 m contenant au moins 15 paléosols. Les paléosols d'Eureka Sound se conforment aux critères appliqués aux sols actuels de type podzolique, gleysolique et organique, et apparaissent au-dessous des sols cryosoliques contemporains. Les paléosols, qui portaient tous une végétation de forêt, se sont développés sur des matériaux fluviatiles et organiques, et leur empilement est le résultat de leur enfouissement par des sédiments fluviatiles. Après leur enfouissement, ils ont été modifiés par des eaux souterraines minéralisantes, par le compactage dû à la charge sédimentaire et glaciaire, et par des failles provoquées par la charge et par les mouvements tectoniques. Les données obtenues durant cette étude facilitent la reconstruction des conditions climatiques, géologiques et pédologiques qui ont régné durant l'Éocène.

INTRODUCTION

The excellent preservation of forest floor and peat material at the fossil forest site on Axel Heiberg Island implies that underlying mineral soils should also be well preserved. Initial descriptions of the fossil forest site at Geodetic Hills (Basinger, 1986; Francis and McMillan, 1987) indicated that exhumed litter layers, essentially surface soil horizons, existed intact and largely unaltered by permineralizing groundwater. Lack of chemical alteration and physical disruption had created a magnificent opportunity to examine complete paleosol profiles from the Eocene.

A paleosol has been defined as "a soil formed in a landscape of the past" (Ruhe, 1956). This simplistic definition has limitations in its application, however. More recently, Fenwick (1985) suggested that the term *paleosol* be confined to soils isolated from present pedogenic processes by reason of burial. Paleosols are best preserved in sedimentary sequences that have accumulated rapidly in a regular episodic fashion over a relatively short time-cycle and in a noneroding manner.

Paleosols have value as paleoenvironmental indicators. They represent the pedological record, which is a reflection of environmental factors that existed during soil formation (Bullock, 1985). Certain morphological properties, including the degree of reddening, nature and distribution of organic matter and the depth of leaching, can be of value when interpreting the climatic conditions under which soil formation took place. This is especially so when a similar set of morphological properties occurs in more than one paleosol in a stacked sequence in a single geological section. This indicates that such paleosols were formed under similar environmental conditions, implying that such conditions existed over a prolonged period.

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[1] Land Resource Research Centre Contribution No. 89-112

[2] Land Resource Research Centre, Research Branch, Agriculture Canada, Ottawa, Ontario K1A 0C6

[3] Land Resource Research Centre, Research Branch, Agriculture Canada, Box 2703, Whitehorse, Yukon Territory Y1A 2C6

Paleogene plant-fossil-bearing beds have been described in the Canadian Arctic (McKenna, 1980; Francis, 1988), in Spitsbergen (Schweitzer, 1980) and even in Antarctica (Kemp and Barrett, 1975), together with their associated floral and faunal assemblages. Although there are likely to be paleosols associated with former peat beds, no detailed pedogenic morphologies have been presented, nor have inferences about soil development been reported. The authors are not aware of any published detailed descriptions of pre-Pleistocene paleosol profiles from the Canadian Arctic.

Tertiary paleosol morphologies have been described in conjunction with the study of sedimentary environments in the Rocky Mountain region of the western United States. Paleosol morphologies in the Willwood Formation of Wyoming indicate the climate of the Paleocene to be temperate, with marked seasonality and mean annual temperatures near 10°C (Kraus, 1988). Lehman (1987) described the development of petrocalcic horizons under semiarid conditions in West Texas, also for Paleocene time.

Significant warming in the Eocene led to the accumulation of organic matter in West Texas soils (Lehman, 1987) and the development of Podzolic soils in Wyoming (Kraus and Brown, 1988). Kraus (1988) reported coniferous forests composed of *Metasequoia* and *Glyptostrobus* in Wyoming's Bighorn Basin and suggested a mean annual temperature of 14°C. This paleoenvironmental scenario is similar to that proposed for the Eocene on Axel Heiberg Island (Francis, 1988).

The mid-continent area appears to have become much drier toward the end of the Eocene. Studies of Upper Eocene and Oligocene fossil soils in South Dakota show evidence of change from woodland to savanna to open grassland (Retallack, 1983). The soils changed from Luvisolic-like soils to highly calcic soils. Dry conditions appear to have been widespread through the Neogene. In the Miocene-Pliocene Ogallala Formation of the Texas Panhandle and eastern New Mexico, 22 calcic paleosols have been identified, suggesting a relatively stable, semiarid environment in the Miocene and Pliocene (Gustavsen and Winkler, 1988).

Paleobotanists have long agreed that Tertiary global climates, particularly during the Eocene, were much warmer than the climate of today, with temperate, broad-leaved deciduous and mixed coniferous forests existing at high latitudes (Wolfe, 1980). What is not clearly understood is the mechanism by which so-called temperate climates could have occurred at high latitudes at that time. Some of the hypotheses for the generation of warmer temperatures are changes in one or more of the following: the obliquity of the Earth's spin axis, the "solar constant," the pattern of global heat circulation, and atmospheric concentrations of greenhouse gases (McKenna, 1980).

The objectives of this study are to identify the range and types of soil development in the various materials in the Geodetic Hills area; to characterize these pedogenic features through detailed profile descriptions and sampling; to use the paleosol morphologies in interpreting paleoclimate; and to reconstruct landscape development of the area during Eocene time. The field methods used and the observations made during the study are summarized here and the paleosols are related to the sedimentological framework established for the area (Ricketts, *this volume*).

THE FOSSIL FOREST SITE

The fossil forest site in the Geodetic Hills area is located near the east coast of Axel Heiberg Island, approximately 60 km west of Eureka, N.W.T. (Fig. 1), at latitude 79°55'N and longitude 88°58'W.

The fossiliferous beds lie within the Upper Coal Member of the Buchanan Lake Formation (Ricketts, *this volume*). This formation, represented by a number of facies associations, is the highest stratigraphic unit of the Cretaceous-Tertiary Eureka Sound Group. In the Geodetic Hills area the formation is estimated to have a thickness of up to 1000 m (Ricketts, 1986). The preserved flora suggest that the beds are of Middle Eocene age (Ricketts and McIntyre, 1986).

The sediments consist of weakly consolidated sands, silts and clays (Ricketts, 1986) deposited during a major period of uplift, faulting and folding in the Eurekan orogeny (Ricketts, 1987). The nonmarine, fine-textured sediments are interbedded with prominent, weakly to moderately lignified coal beds.

The area now has an arctic climate, characterized by long, cold winters and short, cool summers. On the basis of climate normals from the Eureka weather station (Atmospheric Environment Service, 1982), it is estimated that the study area has a mean annual temperature of -20°C and annual precipitation of 65 mm, 70 per cent of which falls as snow. The warmest month of the year is July, with a mean daily temperature of about 5°C; the coldest month is February, with a mean daily temperature of approximately -38°C.

The region is underlain by continuous permafrost. Active layer thickness ranges from 20 to 80 cm, depending on the moisture regime and the texture of the material. Most of the area has little vegetative cover to insulate the ground. Surface soils are thus exposed to continuous sunlight, which, with wind and lack of rain, can result in a marked moisture deficit through evaporation. The surface soils tend to be very dry with many contemporary soils exhibiting surface crusts of salt crystals. The vegetation that does exist consists of a variety of arctic plants that thrive under arid conditions (Basinger, 1986).

Contemporary soils belonging to the Cryosolic order (Agriculture Canada Expert Committee on Soil Survey, 1987) are the only observed surface soils. Turbic Cryosols are the dominant soils in the study area and occur on moderately well to poorly drained, medium to fine textured material. Smaller areas of Static Cryosols occur on well drained and rapidly drained, coarse textured material.

METHODS

Eocene sedimentary strata and associated pedological features were examined in the field. Samples were collected in the steep-sloped sections exposed in hillsides created by Pleistocene and post-Pleistocene fluvial downcutting.

The hillsides were first traversed on foot to obtain an overview of the range of material and the range of soil-like weathering features and organic deposits in the Geodetic Hills area. Exposures were examined along approximately 10 km of traverses around the field camp (Fig. 2). A detailed examination of horizontally continuous individual features was made. Deposits that appeared to have formed at different times, represented by different layers, were examined.

Criteria used to identify paleosols

Distinguishing pedological features from those formed by postburial geochemical (i.e., groundwater) processes was difficult. In addition, fluvial reworking of soil sola, compaction and even tectonic displacements had affected the morphologies (soil features such as soil structure, colour, horizonation) examined. The following criteria were used to establish whether a morphological feature should be considered pedogenic in origin:

a. horizontal continuity for at least the length of a pedon with horizons arranged in a genetically logical sequence (e.g., decreasing degree of weathering with depth)
b. indication of soil surface, specifically: discrete layer(s) of organic matter or the presence of permineralized organic surface horizons
c. gradation in degree of decomposition of organic matter with depth

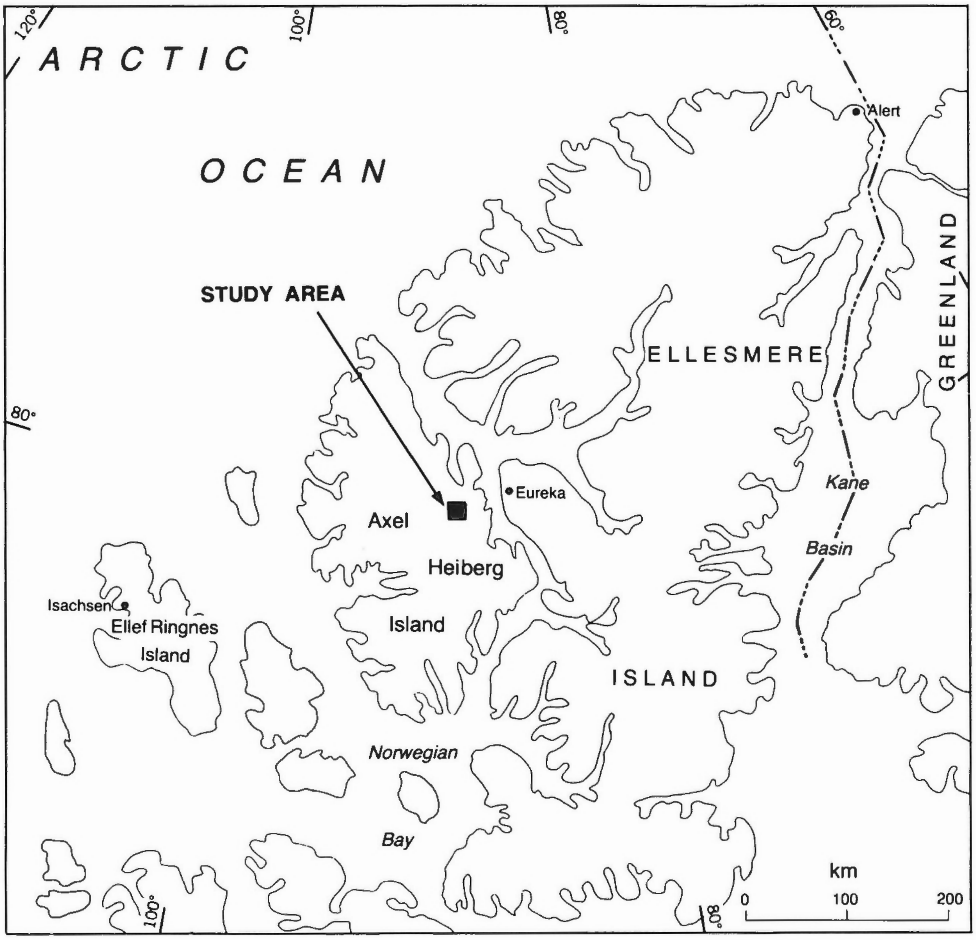

Figure 1. Location of the fossil forest study site in the Geodetic Hills area, Axel Heiberg Island.

d. horizon sequence from purely organic to organo-mineral to mineral horizons
e. presence of ancient roots or root channels
f. evidence of catenary association
g. imposition of pedogenic features over sedimentary features.

Not all of these features are necessarily found at one site; if any of these features are present, it is assumed pedogenic processes were the cause. The paleosols (whether or not truncated) described below as representative soils, in each case met at least two of the criteria.

Description and sampling

The profiles were first traversed both vertically and horizontally, and pedological features were observed. The soil profiles and paleocatenary sequences were then described and sampled (Fig. 2). All sampling and descriptions were of vertical sections. Standard methods of description following Day (1982) were used, and soil horizons were designated according to the Canadian System of Soil Classification (Agriculture Canada Expert Committee on Soil Survey, 1987). Soil samples were collected for analytical characterization in a variety of ways:

a. for individual features, composite samples were collected from throughout the length of the feature (organic layers, so-called "white layers," concretionary horizons)
b. for soil profiles, bulk samples were collected from major genetic horizons across the equivalent distance of one pedon
c. where continuous pedogenic development along a catenary sequence was observed, the continuum was partitioned and individual profiles representing modal drainage conditions were described and sampled
d. in order to examine pedological-sedimentological relationships, a continuous description was made within a facies association. Soils and humus forms were examined over a vertical distance of 22 m within the fin-

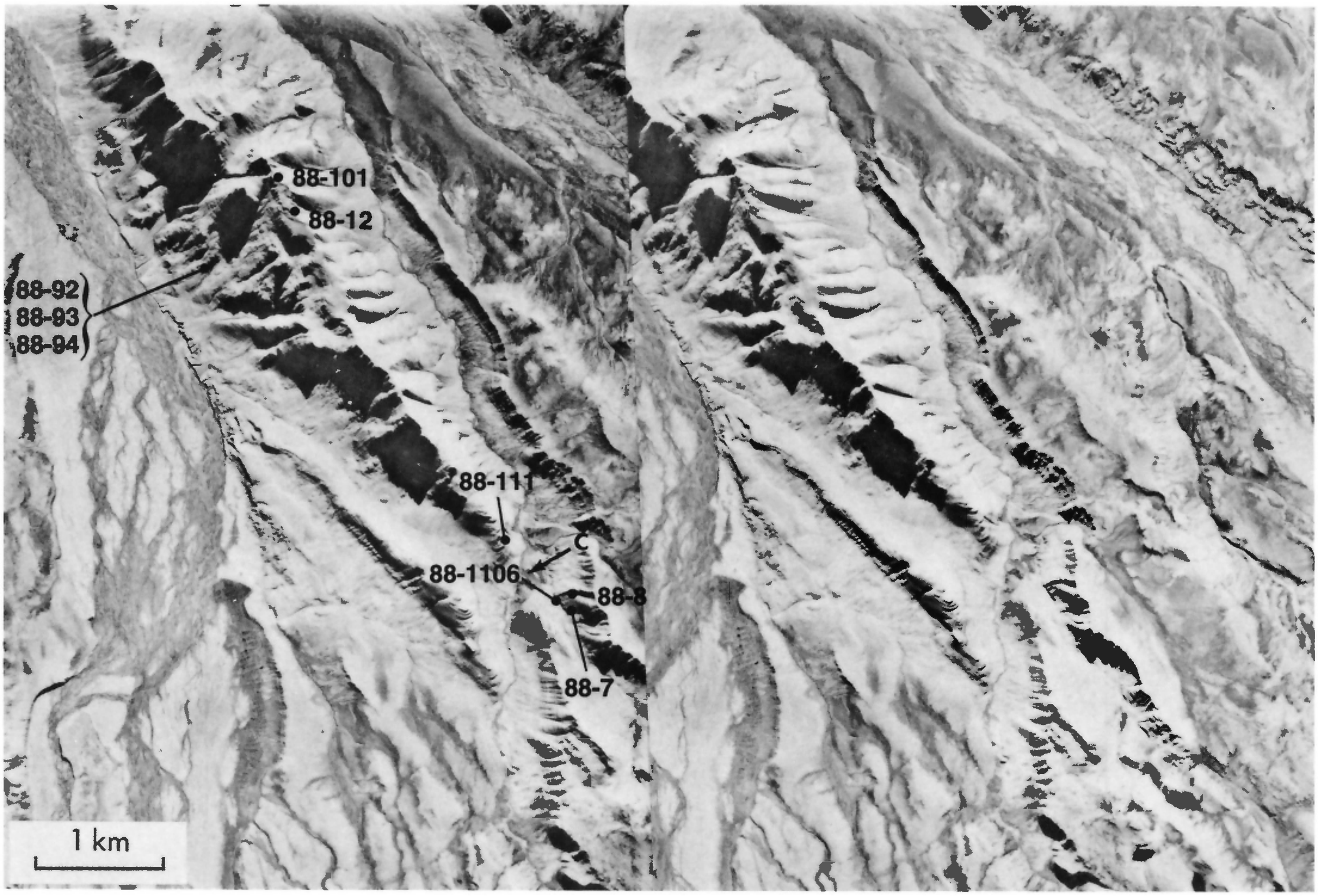

Figure 2. *A stereopair of the fossil forest study area showing the sample sites (sample numbers are given) and the location of the camp (c).*

ing-upward siltstone-coal association, corresponding roughly to elevations 90 to 115 m of Ricketts (*this volume*, Fig. 6).

Additional samples were collected to determine bulk density and to prepare thin sections for future micromorphological study.

Analytical methods

The analytical methods used are outlined in Sheldrick (1984). The methods used are as follows: pH determined by using the 0.01M $CaCl_2$ method; total carbon by using the LECO-600 induction furnace method; bulk density by the core method; particle-size distribution by the pipette method; and the oxalate-extractable Al and Fe by the acid ammonium oxalate extraction method.

RESULTS

During this research it became apparent that different disciplines often use different terms to describe the same sedimentary strata. This is especially so with organic deposits. For example, initial petrographic analyses by geologists indicate that the organic material in this deposit varies from peat to lignite, and that none of the material has reached the bituminous stage (B.D. Ricketts, pers. comm.). In this paper, however, all of these materials are referred to as peat (e.g., lignified woody peat). Similarly, geologists refer to the mineral components as sandstone, siltstone and mudstone. In this paper, these materials, because of their lack of consolidation, are referred to as sands, silts and clays.

Paleosols are classified using the taxonomy applied to contemporary Canadian soils. The chemical criteria for the classification of Podzolic soils, however, are being revised. The new oxalate-extractable Fe and Al criteria proposed for the classification of Podzolic soils by Wang (1989) have been applied in this paper to the classification of the Podzolic paleosols. According to these new criteria, for Podzolic soils the oxalate-extractable Al + $^1/_2$ of the oxalate-extractable Fe should be equal to, or greater than, 0.7 per cent.

Description of paleosols

Paleosols that had developed at various elevations under various paleodrainage conditions within deposits of the Buchanan Lake Formation were examined and sampled in order to determine their characteristics. Well drained and imperfectly drained Podzols and poorly drained Gleysols and Organic paleosols were found. Regosols were also found but were not studied in detail because of the immaturity of this soil type. In a number of cases the Podzols and Gleysols occurred in the same layer of the deposit, clearly indicating the paleodrainage boundaries. The Organic paleosols, on the other hand, formed a separate, continuous layer many kilometres long. All of the paleosols developed in the Buchanan Lake Formation are referred to as Eureka Sound paleosols. The descriptions and characteristics of these paleosols are presented in the following sections.

Podzolic paleosols

The sola of the Podzolic paleosols, which developed on well drained sandy material, are 77 to 140 cm thick (Table 1). The Bf horizons are 65 to 77 cm thick and are frequently associated with cemented orstein and concretionary layers. The colours of the uppermost Bf horizons are 5YR and 7.5YR (Table 1). The leached Ae horizon, when present, is usually very thin or discontinuous. Where surface organic horizons are present, they are also discontinuous and permineralized. A profile description of a Podzolic paleosol is presented in Table 2 and shown in Figure 3, A,B. The sola of these Podzolic paleosols and the underlying parent materials are commonly associated with faulting and cracks resulting both from compaction and shifting of materials and from tectonic movements (Fig. 4).

Two Podzolic paleosol pedons are presented in Table 2. Pedon 88-92 is representative of the Podzolic paleosols found in the lower and middle parts of the Buchanan Lake

TABLE 1

Soil morphology of Eureka Sound mineral paleosols

Paleosol	Pedon No.	Solum thickness (cm)	Thickness of B horizon (cm)	Dominant moist colour	
				Uppermost Bf horizon	Lowermost Bf horizon
Podzol	88-51	110	65	7.5YR 3/2	10YR 4/2.5
	88-21	95	62	5YR 3/2.5	5YR 3/3
	88-91	140	66	5YR 4/6	7.5YR 4/6
	88-92	140	77	5YR 4/6	7.5YR 4/4
	88-101	87	70	7.5YR 4/4	10YR 4/3
Gleyed Podzol	88-94	105	34	5YR 4/1	5YR 4/1
Gleysol	88-93	49	—	—	—

Formation; pedon 88-101 is representative of the Podzolic paleosols found in the upper part of this formation. These paleosols show a weakening of soil development in the upper part of the Buchanan Lake Formation. This is seen most readily in the colour of the B horizon. In the Podzols occurring in the upper part of this formation the B horizons are brownish, but they are reddish brown in the Podzols occurring in the middle and lower parts of the formation. The colours of Podzols occurring in the upper part of the Buchanan Lake Formation resemble those of Brunisolic soils; the oxalate-extractable Al and Fe content, however, positively places them in the Podzolic order (Table 3).

The pH of the Podzolic paleosols in the lower and middle parts of the Buchanan Lake Formation ranges between 4.2 and 5.6. The bulk density ranges from 1.03 to 1.05 g/cm^3 in the Bf horizon and is 1.16 g/cm^3 in the cemented and concretionary Bfcc horizon. The oxalate-extractable Al + $^1/_2$ Fe is 1.79 per cent in the uppermost Bf horizon (Table 3).

The pH of the Podzolic paleosols in the upper part of the Buchanan Lake Formation ranges between 3.5 and 4.9. The oxalate-extractable Al + $^1/_2$ Fe is 0.99 per cent in the uppermost Bf horizon (Table 3).

Gleyed Podzolic paleosols

The sola of the Gleyed Podzolic paleosols, which developed on imperfectly drained loamy sand material, are slightly over one metre thick. The Bf horizon is 34 cm thick and has a colour of 5YR (Table 1). These paleosols have thin surface organic horizons underlain by thick Ah horizons. The lighter

TABLE 2

Descriptions of Podzolic paleosol pedons

Horizon	Depth (cm)	Description
Podzolic paleosol 88-92		
Fu*	0 - 15	Dark reddish brown (5YR 2.5/2 m); permineralized wood and other forest litter material; clear, smooth horizon boundary; 10 to 15 cm thick.
Ae	15 - 22	Light brownish gray (2.5Y 6/2 m); sand; single grain structure; nonsticky, loose, nonplastic consistence; clear, broken horizon boundary; discontinuous; 0 to 7 cm thick.
Bf1	22 - 55	Yellowish red (5YR 4/6 m); sand; weak, fine to medium, granular structure; nonsticky, loose, nonplastic consistence; gradual, smooth horizon boundary; 20 to 33 cm thick.
Bf2	55 - 88	Yellowish red (5YR 4/6 m); sand; weak, fine to medium, granular structure; nonsticky, loose, nonplastic consistence; gradual, smooth horizon boundary; 25 to 33 cm thick.
Bfcc	88 - 99	Dark brown (7.5YR 4/4 m); sand; weak, fine to medium, granular structure; nonsticky, loose, nonplastic consistence; clear, smooth horizon boundary; moderately cemented; medium size iron-manganese concretions; 8 to 11 cm thick.
BCcc1	99 - 118	Grayish brown (10YR 5/2 m); sand; single grain structure; nonsticky, very friable, nonplastic consistence; gradual, smooth horizon boundary; moderately cemented; medium size iron-manganese concretions; 15 to 19 cm thick.
BCcc2	118 - 140	Grayish brown (10YR 5/2 m); sand; single grain structure; nonsticky, loose, nonplastic consistence; gradual, smooth horizon boundary; moderately cemented; medium size iron-manganese concretions; 15 to 22 cm thick.
C	140 - 175	Dark gray (10YR 4/1 m); sand; single grain structure; nonsticky, loose, nonplastic consistence.
Podzolic paleosol 88-101		
Fu	0 - 5	Dark reddish brown (5YR 2.5/2 m); permineralized wood and other forest litter material; clear, smooth horizon boundary; 3 to 5 cm thick.
Ae	5 - 17	Grayish brown (10YR 5/2 m); sand; structureless, single grain structure; nonsticky, loose, soft, nonplastic consistence; abrupt, wavy horizon boundary; 8 to 12 cm thick.
Bf1	17 - 39	Strong brown (7.5YR 4/4 m); sand; structureless, single grain structure; nonsticky, very friable, slightly hard, nonplastic consistence; clear, wavy horizon boundary; 15 to 22 cm thick.
Bf2	39 - 87	Dark brown (10YR 4/3 m); sand; structureless, single grain structure; nonsticky, loose, nonplastic consistence; gradual, smooth horizon boundary; 35 to 48 cm thick.
C	87 - 115	Dark grayish brown (2.5Y 4/2 m); sand; structureless, single grain structure; nonsticky, loose, nonplastic consistence.

Fu*, permineralized organic horizon

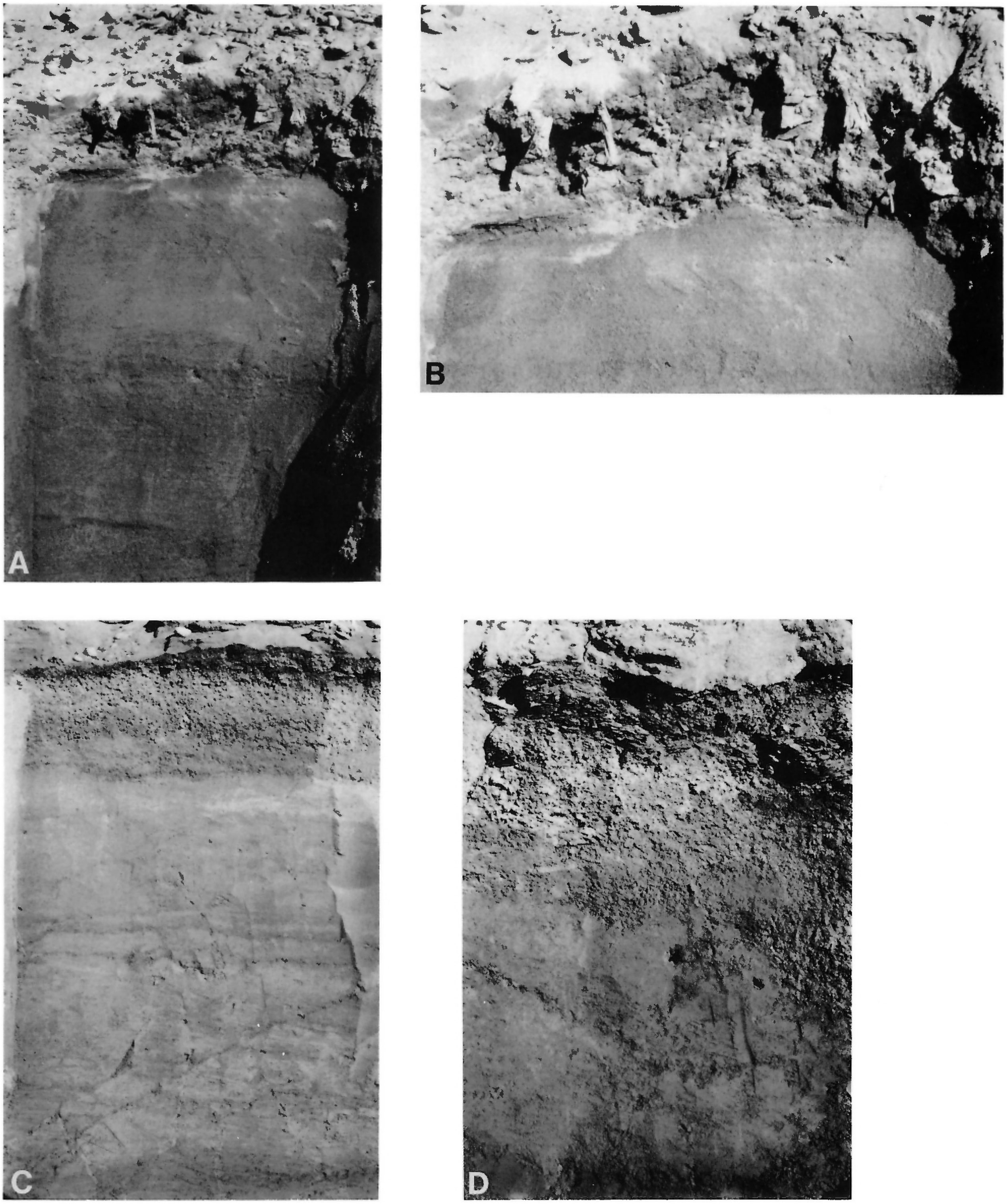

Figure 3. *Eureka Sound mineral paleosols: A, well drained Podzolic paleosol (Table 2, pedon 88-92); B, surface horizon of the well drained Podzolic paleosol showing the permineralized organic horizon (Table 2, pedon 88-92); C, imperfectly drained Podzolic paleosol (Table 4, pedon 88-94); and D, poorly drained Gleysolic paleosol (Table 4, pedon 88-93). All of these paleosols are from the Geodetic Hills area, Axel Heiberg Island.*

colour of the lower part of these Ah horizons indicates that some leaching has occurred. Coarse, distinct to prominent mottles commonly occur in the lower Bf horizons as well as in the BC and C horizons. A profile description of a Gleyed Podzolic paleosol is presented in Table 4 and shown in Figure 3C.

The pH of these paleosols ranges between 5.1 and 6.5. The bulk density of the Bf horizon is 1.38 g/cm³, and for the Cg horizon it is 1.26 g/cm³. The oxalate-extractable Al + $^1/_2$ Fe is 0.76 per cent in the uppermost Bf horizon (Table 3).

Gleysolic paleosols

The sola of the Gleysolic paleosols, which developed on poorly drained loamy sand material, are slightly less than 50 cm thick (Table 1). These paleosols have organic-rich mineral surface horizons (Ah) or well preserved humic to fibrous organic horizons. The Ah horizon, if present, shows signs of leaching. These paleosols have grayish[4] colours indicative of reducing conditions. Medium and coarse mottles occur in all mineral horizons. A profile description of a Gleysolic paleosol is presented in Table 4 and shown in Figure 3D.

The pH of these paleosols ranges between 3.5 and 5.6. The bulk density of the Aheg horizons is 1.28 g/cm³, and for the Cg horizon it is 1.29 g/cm³ (Table 3).

Organic paleosols

The Organic paleosols developed under poorly to very poorly drained conditions with shallow surface waters. They are composed of fibrous woody peat. All coarse material (twigs, cones, branches and logs) in these peats is severely

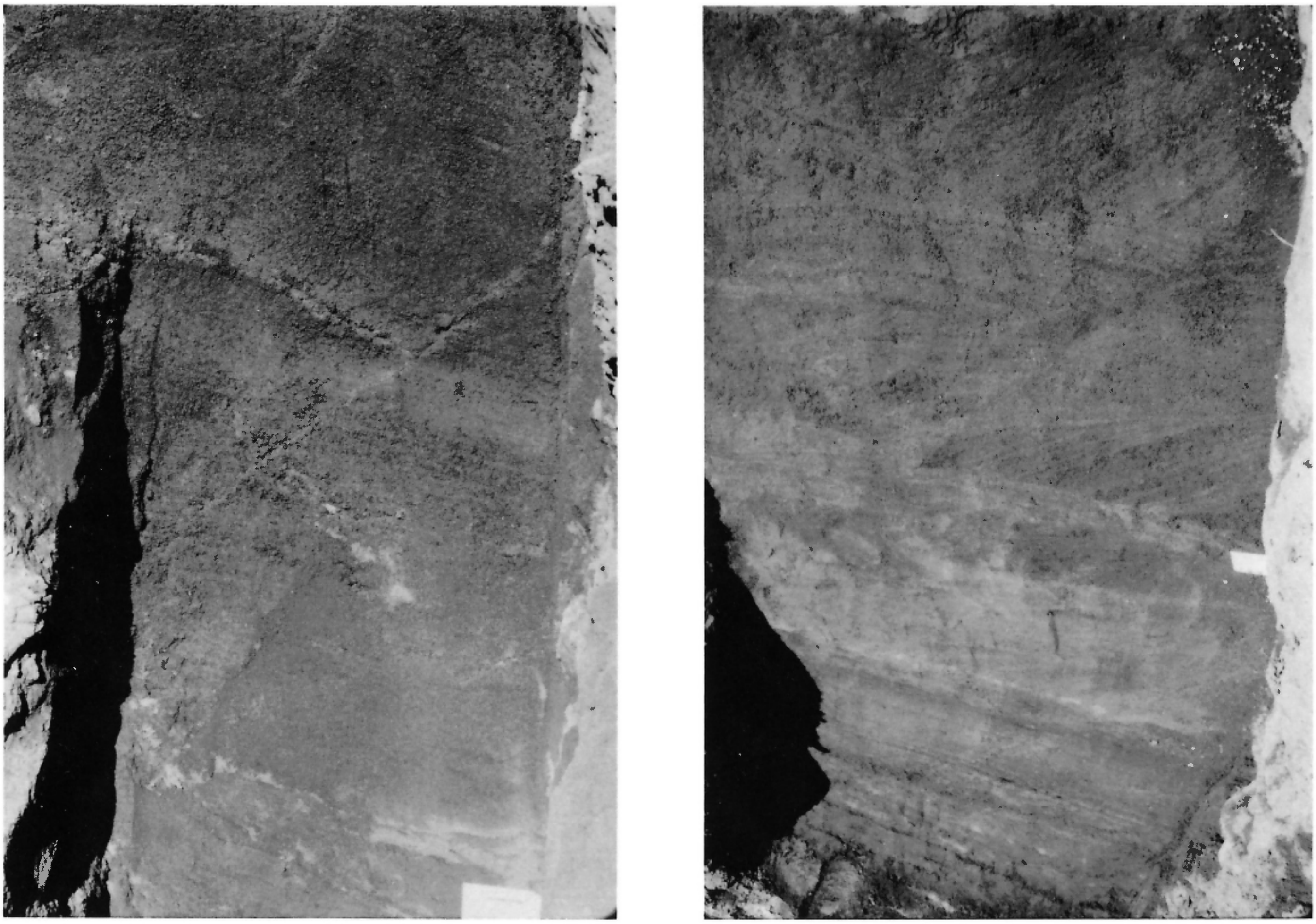

***Figure 4**. Faults and cracks associated with the Eureka Sound paleosols. Geodetic Hills area, Axel Heiberg Island. The photographs show an approximately 1 m vertical portion of the pedon.*

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[4] Spelling to conform with Canadian System of Soil Classification.

flattened because of the pressure resulting from the weight of the overlying sediments and Pleistocene glacial ice. The wood in this peat material has not been permineralized, but rather it has been preserved in a mummified and slightly lignified state. The thickness of this compressed peat is generally 50 to 80 cm, but may be as much as 200 cm. All of these organic soils contain nodules of amber up to 4 cm in size. A profile description of an Organic paleosol is presented in Table 5 and shown in Figure 5.

The pH of these paleosols ranges between 3.4 and 5.2. The bulk density of the Om horizon is 0.72 g/cm^3 (Table 3).

Pedological-sedimentological relationships

Soil development within the fining-upward siltstone-coal association

Paleosols within this facies association developed in a moist floodplain environment, subject to frequent inundation

TABLE 3

Analytical data for Eureka Sound paleosols

Horizon	pH	Total	Oxalate Extractable		Bulk density	Particle size distribution (% <2 mm)		
	(CaCl$_2$)	C%	Fe%	Al%	g/cm^3	Total sand	Silt	Clay
Podzolic paleosol 88-92								
Fu	5.6	1.44	9.79	0.24	–	66.7	26.3	6.9
Ae	4.2	0.44	2.41	0.14	–	87.7	7.3	2.9
Bf1	4.7	0.31	3.27	0.16	1.03	92.3	3.5	4.2
Bf2	4.6	0.15	1.94	0.15	1.05	94.7	2.6	2.7
Bfcc	4.5	0.46	2.35	0.16	1.16	93.0	3.9	3.1
BCcc1	5.2	0.47	2.22	0.16	–	93.1	4.5	2.7
BCcc2	5.2	0.42	2.06	0.16	–	91.3	6.8	1.8
C	5.1	0.51	2.44	0.15	–	92.8	5.1	2.0
Podzolic paleosol 88-101								
Fu	–	–	–	–	–	–	–	–
Ae	4.5	0.37	0.79	0.04	–	86.8	11.6	1.6
Bf1	4.9	0.37	1.89	0.05	–	91.8	6.0	2.2
Bf2	3.5	0.38	1.89	0.05	–	91.8	6.0	2.2
C	5.0	0.50	1.42	0.06	–	90.1	7.6	2.2
Gleyed Podzolic paleosol 88-94								
FH	5.6	12.59	0.95	0.07	–	7.2	68.1	24.6
Ah	5.3	1.48	0.65	0.06	1.03	13.6	64.5	21.8
Ahe	5.1	0.62	1.32	0.07	1.30	19.4	71.9	8.5
Bf	5.1	0.68	1.37	0.08	1.38	60.1	37.7	2.1
Bfg	6.2	0.70	1.05	0.07	–	82.0	16.7	1.2
BCg	6.5	0.69	2.06	0.12	–	80.6	18.5	0.8
Cg	6.1	0.90	1.17	0.11	1.26	84.8	14.7	0.4
Gleysolic paleosol 88-93								
Om	3.5	27.38	5.52	0.09	–	–	–	–
Aheg1	4.8	2.61	0.86	0.04	1.28	22.3	65.0	12.6
Aheg2	5.6	0.69	0.75	0.05	1.28	61.7	35.5	2.7
Cg	5.8	0.64	0.60	0.06	1.29	71.2	26.8	1.9
Organic paleosol 88-111								
Om	3.4	33.76	1.79	0.24	0.72	–	–	–
Ahg	5.2	2.28	0.37	0.04	–	24.5	59.1	16.2
Cg1	5.0	1.63	1.16	0.07	–	41.6	50.0	8.2
Cg2	5.5	1.42	2.20	0.10	–	44.9	44.1	10.8

and deposition. Organic (woody peat), Gleysolic and Regosolic soils, as shown in Figures 3C and 4 and Tables 4 and 5, are characteristic. The lengths of time during which the soils formed varied by an order of magnitude. Estimates based on both the thickness of the organic layer and contemporary soil development models indicate that the Regosols, the most immature of the soils, likely developed over a period ranging up to 10^2 years and that the Gleysols and Organic soils likely developed over a period ranging from 10^3 to 10^4 years.

The floodplain environment evidently was favourable for the accumulation of organic matter on the soil surface; a variety of humus forms and peat material can be seen within the facies. Most prominent is the woody peat, all of which is compressed. Compressed, intermixed mineral material and woody peat also occurs. A second humus form consists of highly decomposed, amorphous humus. This latter is variously compressed and apparently uncompressed and friable. A third humus form is composed of comminuted organic fragments that appear to be reworked.

Similar humus forms in a permineralized state are also evident. Litter material incorporating leaf and needle imprints and root channels, and woody material, as well as massive ferrilite blocks with and without evidence of root channels are present. All permineralized humus forms are shattered.

A trench was dug to expose 22 m of vertical relief in the fining-upward siltstone-coal association of the Buchanan Lake Formation (Ricketts, *this volume*) (Fig. 6). In this section fifteen discrete paleosols were found: four Organic paleosols dominated by woody peat and eleven mineral paleosols with humus-rich horizons. Since only a small fraction of the total thickness of the association was sampled, it is very likely that the whole association has many more paleosols than were observed in the 22 m section.

The particle size distribution in the samples collected in this section indicates that the clay content of the sediments varies greatly. The clay content of some of the layers is low (<10%), indicative of an active alluvial environment, while

TABLE 4

Descriptions of Gleyed Podzolic and Gleysolic paleosol pedons

Horizon	Depth (cm)	Description
Gleyed Podzolic paleosol 88-94		
FH	0 - 4	Very dark gray (2.5Y 3/0 m); forest litter, composed dominantly of woody materials; moderately decomposed; strong, coarse platy; gradual, smooth horizon boundary; 3 to 4 cm thick.
Ah	4 - 15	Very dark gray (7.5YR 3.5/0 m); silt loam; strong, medium to coarse, subangular blocky structure; nonsticky, very friable, hard, nonplastic consistence; clear, smooth horizon boundary; 8 to 11 cm thick.
Ahe	15 - 32	Dark gray (7.5YR 4/0 m); silt loam; moderate to strong, medium, granular structure; nonsticky, very friable, slightly hard, nonplastic consistence; clear, smooth horizon boundary; 10 to 17 cm thick.
Bf	32 - 43	Dark gray (5YR 4/1 m); sandy loam; very weak, fine, granular structure; nonsticky, loose, nonplastic consistence; gradual, smooth horizon boundary; 7 to 11 cm thick.
Bfg	43 - 66	Dark gray (5YR 4/1 m); loamy sand; very weak, fine, granular structure; nonsticky, loose, nonplastic consistence; common, coarse, distinct mottles (10YR 4/4); gradual, wavy horizon boundary; 18 to 23 cm thick.
BCg	66 - 105	Very dark gray (5YR 3/1.5 m); loamy sand; weak, fine, granular structure; nonsticky, loose, nonplastic consistence; common, coarse, distinct mottles 10YR 5/5); gradual, wavy horizon boundary; 30 to 39 cm thick.
Cg	105 - 170	Dark gray (2.5Y 4/0 m); loamy sand; very weak, subangular blocky structure; nonsticky, loose, nonplastic consistence; common, coarse, distinct mottles (10YR 5/5).
Gleysolic paleosol 88-93		
Om	0 - 13	Black (2.5Y 2/0 m); woody peat, composed dominantly of woody materials; moderately decomposed; strong, coarse platy; clear, wavy horizon boundary; 10 to 13 cm thick.
Aheg1	13 - 35	Dark gray (5YR 4/1 m); sandy loam; weak to moderate, medium, platy structure; nonsticky, very friable, slightly hard, nonplastic consistence; common, prominent, medium mottles (10YR 4/4); gradual, smooth horizon boundary; 17 to 22 cm thick.
Aheg2	35 - 49	Very dark gray (7.5YR 3.5/0 m); loamy sand; weak to moderate, medium, subangular blocky structure; nonsticky, very friable, slightly hard, nonplastic consistence; common, prominent, coarse mottles (10YR 4/4); gradual, smooth horizon boundary; 9 to 14 cm thick.
Cg	49 - 110	Dark gray (5YR 4/1 m); loamy sand; structureless, single grain structure; nonsticky, loose, nonplastic consistence; common, prominent, coarse mottles (10YR 4/4).

that of other layers is very high (>40%), indicative of a lacustrine-like environment (Fig. 6). Peat layers are usually underlain by fine-textured sediments.

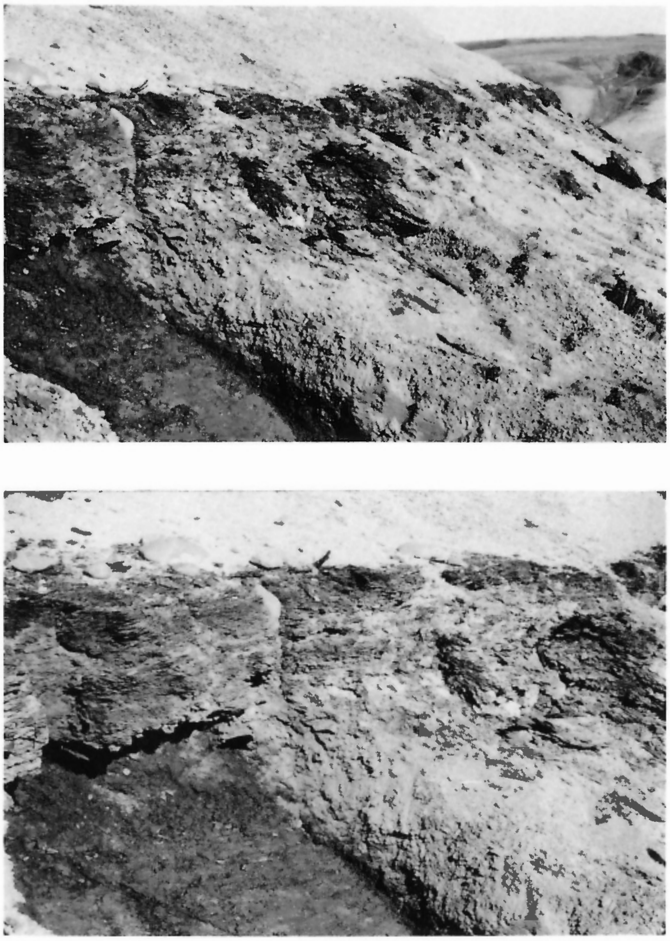

Figure 5. Organic paleosols showing the lignified woody peat material and the underlying gleyed mineral material (Table 5, pedon 88-111).

Soil development within the fining-upward sandstone-coal association

Paleosols within this facies association probably developed in slightly elevated interchannel areas (levees) that were not subject to frequent inundation. Well drained to imperfectly drained Podzolic soils developed on sandy material are characteristic (Table 2, Fig. 3A, C). The degree of pedogenic development and the solum thickness indicate that these areas had likely been stable for a considerable period of time (10^3 to $>10^4$ years) before burial.

Many of the permeable sandy materials appear to have been affected by mineralization and are cemented with iron and silica. A certain amount of reddish staining is also evident. In many cases it is difficult to differentiate pedogenic influence from geochemical alteration. Litter material associated with well drained soil surfaces is also very often permineralized in many places.

Soil development within the thick sandstone association

This facies association is present at the top of the Buchanan Lake Formation in the Geodetic Hills area. Although it is without significant coal beds, humic peat and Podzolic paleosols with brownish B horizons can be found (Pedon 88-101, Table 2). These Podzolic paleosols probably formed as a result of either less intense weathering conditions (cooler, drier) or shorter periods of weathering than those for the underlying associations.

The paleosols and associated organic matter are generally uncompacted and not permineralized. The humus is friable and well decomposed.

TABLE 5

Description of an Organic paleosol pedon

Horizon	Depth (cm)	Description
Organic paleosol 88-111		
Om	0 - 65	Black (2.5Y 2/0 m); lignified woody peat, composed dominantly of leaves, needles, twigs and logs; moderately decomposed; moderately coarse platy; clear, smooth horizon boundary; 60 to 65 cm thick.
Ahg	65 - 91	Very dark gray (2.5Y 3/0 m); silt loam; moderately strong, medium to coarse, subangular blocky structure; slightly sticky, friable, slightly hard, slightly plastic consistence; common, fine, prominent mottles (7.5YR 4/4); clear, smooth horizon boundary; 20 to 26 cm thick.
Cgl	91 - 112	Dark gray (10YR 4/1 m); silt loam; moderate to strong, medium to coarse, subangular blocky structure; slightly sticky, firm, hard, slightly plastic consistence; common, fine, prominent mottles (7.5YR 4/4); gradual, smooth horizon boundary; 15 to 21 cm thick.
Cg2	112 - 165	Dark gray (10YR 4/1 m); loam; moderate to strong, medium to coarse, subangular blocky structure; slightly sticky, firm, hard, slightly plastic consistence.

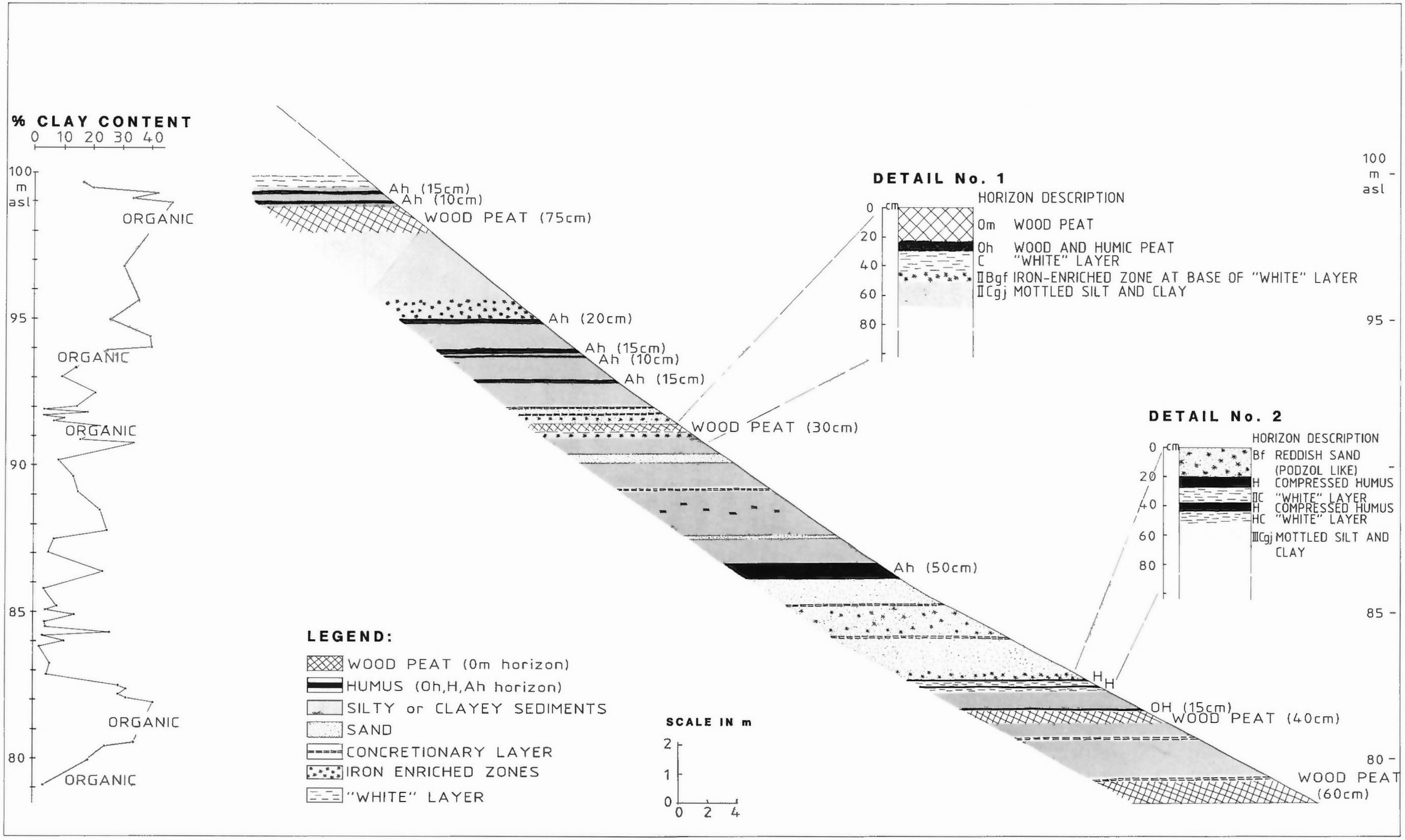

***Figure 6**. Paleosols developed along a 22 m section in the fining-upward coal member of the Buchanan Lake Formation.*

DISCUSSION

The following reconstruction of the paleoenvironment, including climate and pedological events, during the Eocene is based on 1) field observations made at the fossil forest site on Axel Heiberg Island; 2) chemical, physical and morphological data relating to the Eureka Sound paleosols obtained during this study; and 3) present-day models for contemporary soil development.

Soil landscape environment

The Eocene landscape of the fossil forest area was a dynamic sedimentary lowland composed of active and abandoned river channels, lakes and a large area of low-lying swamp-type wetlands (Fig. 7). The river channels meandered freely across this flat, sandy, fluvial landscape. The areas adjacent to these river channels (levees) were slightly elevated above the swampy interchannel areas. During high water levels the sediment-rich river waters flooded the low-lying areas and provided sandy and silty sediments that eventually filled the shallow basins. The positions of the river channels continually changed because of sedimentation and erosion. As a result of these processes the landscape was very dynamic. This is well demonstrated by the trench data (Fig. 6). In the 22-m section, at least 15 paleosols with well developed organic horizons were found. The development of the organic horizons indicates a long-lived stable environment without fluvial deposition. The varying thicknesses of the organic horizons (5 to 75 cm) indicate that the length of time between sedimentary inundations also varied. In addition, the presence of only one thick sedimentary layer (less than 3 m thick) in the section indicates that there was probably only one period of major flooding. Paleosols found in the 22-m section of this study display soil sequences and deposition patterns similar to those commonly found in contemporary alluvial deposits.

The soil that developed on each sedimentary surface varied, depending on the water regime (drainage) (Fig. 7). Podzolic soils developed on the slightly higher levees along active and abandoned river channels, whereas Gleyed Podzolic soils developed on imperfectly drained sites. The development (maturity) and thickness of these paleosols indicate that these areas were stable for a long period of time.

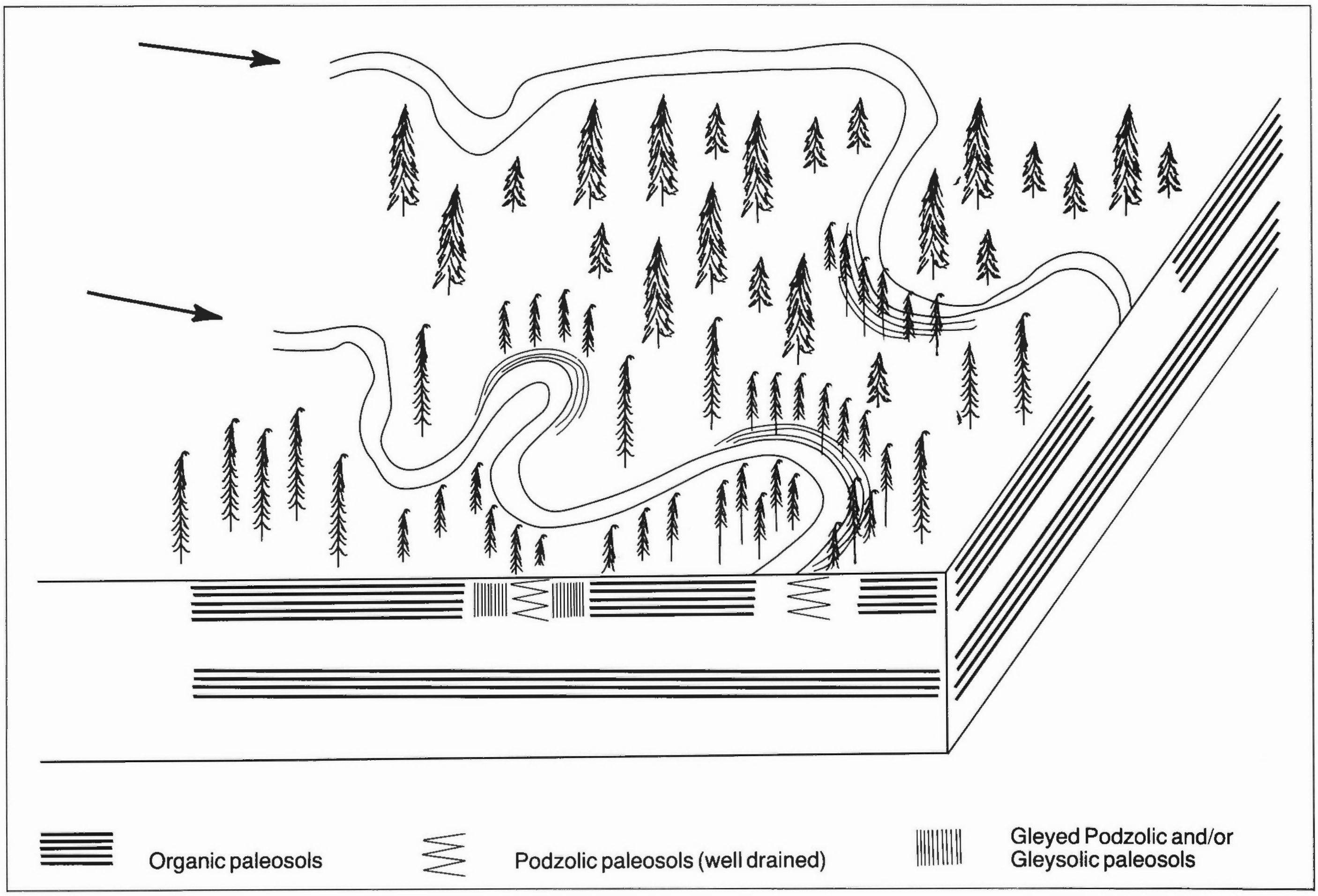

***Figure** 7. Diagrammatic reconstruction of the Eocene landscape of the fossil forest study area.*

The low-lying areas, which may have dominated this Eocene landscape (based on the extensive peat deposits), were poorly or very poorly drained. Gleysolic paleosols developed on poorly drained depressions and Organic paleosols developed on very poorly drained areas (Fig. 7). Some Gleysolic and Organic paleosols had a sequence of thin peat, silt and sand layers, suggesting that they were subject to frequent, periodic floods and deposition. Most of the Organic paleosols, however, were associated with swamps with high water table and thick deposits of woody peat material (forest swamp vegetation). The forest swamps, which were dominated by dawn redwood (*Metasequoia* sp.) (Basinger, 1986), were productive wetland environments. The high productivity was due to both a warm, moist climate and a high nutrient status. The high nutrient levels were maintained by the rivers, which provided nutrient-rich waters, especially during the high water stage. As a result of this highly productive environment, thick deposits of woody peat developed. The thickest compacted peat deposit measured in the fossil forest area was 2 m deep. In its original, uncompacted form this peat deposit would have been much thicker. The original thickness of this deposit indicates that at least some of the swamps were stable for many thousands of years, until a major flood, with its sediments, buried them and changed the landscape.

White layers

White layers occur throughout the exposed sections of the Buchanan Lake Formation. These layers are up to 50 cm thick, but are usually 5 to 20 cm thick. The white layers are found in a number of situations: between loamy mineral material; with organic material above and gray, reduced, silty mineral material below; or sandwiched between two distinct organic layers. None of the white layers examined were underlain by a well drained Podzolic Bf horizon. Coarse mottles occurred in some of the thicker white layers, indicating that these layers had been exposed to wet conditions. Seven white layers ranging in thickness from 5 to 20 cm were encountered in the 22-m section (Fig. 6). The white layers are described further by Tarnocai et al. (*this volume*).

Soil development

Eureka Sound Podzolic paleosols are similar in morphology and chemical composition to contemporary Podzols that have formed in temperate climates with high or moderately high precipitation and coniferous forest vegetation. In contemporary Podzols, under such conditions, organic acids are produced as a result of the decomposition of organic materials. This, coupled with active mineral weathering, releases Fe, Al and other elements from the parent material, a consequence of which is the combination of fulvic acids with free Fe and Al. The products are translocated, accumulating in the Bf horizon. Although there are significant concentrations of both Fe and Al in the Bf horizons of contemporary Podzols, in the Bf horizons of the Podzolic paleosols there is a high concentration of Fe but practically no Al accumulation. Similar pedogenic processes occur in the Gleyed Podzolic paleosols. Because of the high water table associated with these soils, however, reduction and mottling also occur.

After burial in their intact or truncated forms, the Podzolic paleosols were evidently subjected to permineralization and faulting, which resulted in the additional soil features now observable in these paleosols (Fig. 8). High levels of Fe and Si in mineral-rich groundwater then led to the formation of concretions and caused the permineralization of woody materials in the surface organic horizons. In addition, faulting occurred, possibly as a result of tectonic movement or of settlement due to sediment deposition and Pleistocene glacial ice. This faulting led to displacement of soil horizons along fault lines by as much as 5 to 10 cm.

The Gleysolic and Organic paleosols have continuous surface organic horizons. The difference between these two paleosols is that the Gleysolic pedons are dominated by mineral material while the Organic pedons are dominated by organic (peat) material. The mineral horizons in both of these paleosols are strongly reduced and have gray colours and prominent mottles.

After burial, the Organic paleosols were compacted by the weight of the overlying sediments and Pleistocene glacial ice (Fig. 9). The most visible evidence of this compaction is the numerous severely flattened twigs, branches and logs to be found in these deposits and in the contorted growth rings of the wood structure. Further evidence of this compaction is to be seen in the slight lignification of the woody peat material. Very little permineralization occurred in these organic deposits, however, probably because these paleosols were well sealed by the organic materials and fine textured deposits. In addition, no faulting was found in these Organic paleosols, likely because of the matted structure resulting from deposition of layers of plant material.

Soil-climate relationships

The large basin swamps and associated Organic paleosols that occurred in the fossil forest area during the Eocene epoch are suggestive of a climate that was warm, moist and productive. The high woody component and the large diameter stumps found in these peats indicate the presence of a heavily forested swamp wetland environment similar to that now occurring in Florida and South Carolina.

The use of models derived from the development of contemporary Podzols makes it possible to reconstruct the Eocene environment at the time the Podzolic paleosols developed.

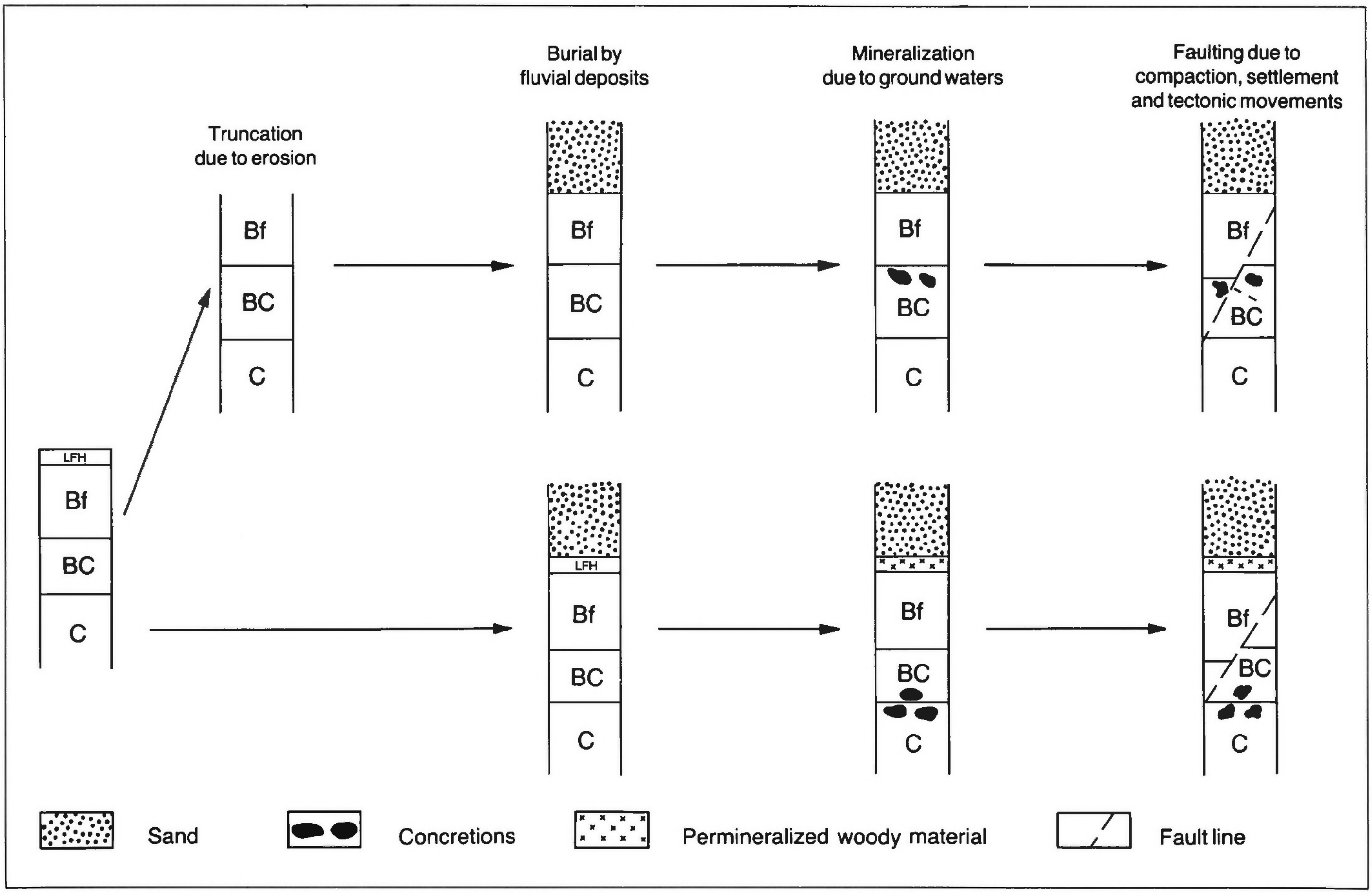

***Figure 8**. Processes affecting mineral paleosols.*

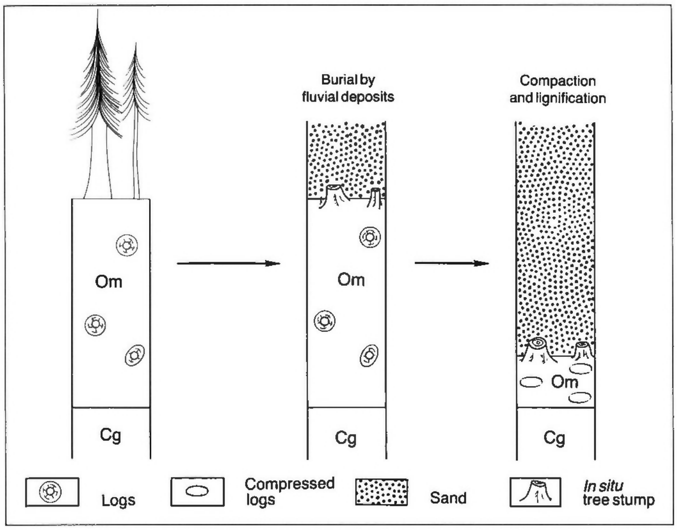

***Figure 9**. Compaction and lignification processes affecting Organic paleosols after burial.*

Contemporary Podzolic soils, similar to the Podzolic paleosols found in the fossil forest area, developed in temperate climates with high or moderately high precipitation. Such conditions now occur along the south coast of British Columbia and the coastal areas of the states of Washington and Oregon. In these climates the temperature seldom drops below 0°C and, when frost occurs, it affects only the surface of the soil.

The Eocene environment associated with the Eureka Sound paleosols was probably characterized by a climate that was temperate to warm temperate with mild, wet winters and warm, moist summers. The vegetation associated with these paleosols was coniferous forest on well drained to imperfectly drained areas and forested basin swamps on poorly and very poorly drained areas.

SUMMARY

1. The Eureka Sound paleosols developed in fluvial and organic material in the Buchanan Lake Formation during the Eocene. These Podzolic, Gleysolic and Organic paleosols are stacked sequentially as a result of repeated burial by fluvial deposition (overbank or crevasse splay deposits).

2. The Podzolic paleosols have reddish brown (5YR and 7.5YR) Bf horizons, which are often cemented and contain iron concretions. If organic surface horizons are present, they are strongly permineralized.

3. The Gleysolic paleosols have strongly gleyed sola with prominent mottles resulting from water-saturated conditions. These paleosols are commonly associated with organic surface horizons that are not permineralized.

4. The Organic paleosols developed on woody peat material derived from forest swamp vegetation. After burial this material was compressed and slightly lignified. These compressed and lignified peat deposits are up to 2 m thick.

5. After burial the Eureka Sound Podzolic paleosols were subjected to mineralization resulting from groundwater, compaction resulting from the weight of overlying sediments and Pleistocene glacial ice, and faulting resulting from loading and tectonic movement.

6. On the basis of contemporary soil development models, our knowledge of the development of these paleosols suggests that the climate in the fossil forest area during the Eocene epoch was temperate to warm temperate with mild, wet winters and warm, moist summers.

ACKNOWLEDGMENTS

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CHARACTERISTICS AND POSSIBLE ORIGIN OF THE WHITE LAYERS FOUND IN THE FOSSIL FOREST DEPOSITS, AXEL HEIBERG ISLAND[1]

C. Tarnocai[2], H. Kodama[2], and C. Fox[2]

Tarnocai, C. et al., 1991. Characteristics and possible origin of the white layers found in the fossil forests deposits, Axel Heiberg Island. In Tertiary Fossil Forests of the Geodetic Hills, Axel Heiberg Island, Arctic Archipelago, R.L. Christie and N.J. McMillan (eds.); Geological Survey of Canada, Bulletin 403, p. 189-200.

Abstract

White layers 5 to 50 cm thick occur throughout the Buchanan Lake Formation in the Geodetic Hills area of Axel Heiberg Island. The particle size distribution curves of the white layer material show polymodal, not normal, distribution. The white layer material contains high amounts of quartz, most of which has a platy morphology, unlike the prismatic morphology of the commonly occurring quartz. In addition, the white layer material contains rod-like and other irregularly shaped particles that resemble plant phytoliths. The lack of B soil horizons underlying these white layers and the platy morphology of the quartz suggest that these layers were not formed by pedological processes. The crystal morphology of the quartz, the unusual particle size distribution, and the rod-like features suggest a biological origin (opal phytolith) for the white layer material. After burial, the amorphous opal phytoliths composing the white layer material underwent a dehydration process, which converted them to crystalline quartz. As a result of this conversion, nearly all original phytolithic morphology was lost, except for those features originally produced in crystalline form by the plants.

Résumé

Dans toute la formation de Buchanan Lake, dans la région des collines Geodetic de l'île Axel Heiberg, apparaissent des couches blanches de 5 à 50 cm d'épaisseur. Les courbes de distribution granulométrique établies pour le matériau stratifié blanc montrent une distribution polymodale, non une distribution normale. Le matériau stratifié blanc contient de fortes proportions de quartz, lequel se caractérise le plus souvent par une morphologie tabulaire, contrairement au quartz habituel, qui présente une morphologie prismatique. En outre, le matériau stratifié blanc contient des particules en formes de bâtonnet et d'autres particules de forme irrégulière qui ressemblent à des phytolites. L'absence d'horizons B du sol au-dessous de ces couches blanches, et la morphologie tabulaire du quartz, suggèrent que ces couches n'ont pas été formées par des processus pédologiques. La morphologie cristalline du quartz, la distribution granulométrique inhabituelle, et les structures en bâtonnets, suggèrent une origine biologique (phytolite opaline) dans le cas du matériau stratifié blanc. Après son enfouissement, le phytolite opaline amorphe composant le matériau stratifié blanc a subi une déshydratation qui l'a converti en quartz cristallin. En raison de cette conversion, presque toute la morphologie initiale du phytolite a disparu, excepté les structures constituées à l'origine sous forme cristalline par les végétaux.

INTRODUCTION

The first report of white layers in the fossil forest deposits of Axel Heiberg Island was made by Francis and McMillan (1987). They referred to these light gray[3] to almost white layers as leached soil horizons. This material could also possibly have been of volcanic, eolian, or fluvial origin. Ricketts (pers. comm., 1988) examined this material but found no evidence of shards, minimizing the possibility of a volcanic origin. He did, however, find that the material contained large amounts of silica and some montmorillonite and kaolinite.

Similar white layers were also reported (Riquier, 1960) in contemporary soils found in East Africa, on the Isle of Réunion. Although Riquier (1960) reported that these white layers, which were 5 to 30 cm thick, had earlier been interpreted as leached soil horizons, he found that they consisted entirely of opal phytoliths (biological silica). The area where the white layers occur has an annual rainfall of 2000 to 4000 mm and mean annual temperatures of 10° to 15°C. The vegetation now growing on these areas is composed of *Acacia heterophylla*, *Asplenium*, *Peteris* and bamboo, and as Riquier (1960) stated, this type of vegetation was the source of the phytoliths.

This paper presents characteristics of the white layer material from the fossil forest site and suggests a possible origin for it.

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[1] Land Resource Research Centre Contribution No. 89-111.

[2] Land Resource Research Centre, Research Branch, Agriculture Canada, Ottawa, Ontario K1A 0C6

[3] Spelling to conform with Canadian System of Soil Classification

DESCRIPTION OF THE AREA

The fossil forest site (latitude 79°55'N, longitude 88°58'W) in the Geodetic Hills area is on the east side of Axel Heiberg Island, approximately 40 km from the coast (Fig. 1). The site is approximately 60 km west of Eureka, Northwest Territories.

The sediments in the study area consist of weakly consolidated sands, silts and clays of the Buchanan Lake Formation (Ricketts, 1986). The Buchanan Lake Formation is the highest stratigraphic unit in the Cretaceous-Tertiary Eureka Sound Group. In the Geodetic Hills area these nonmarine, fine-textured sediments are interbedded with prominent, weakly to moderately lignified coal beds. Sediments were deposited during a major period of uplift, faulting and folding in the Eurekan orogeny (Ricketts, 1987). In the Geodetic Hills, the formation is estimated to have a thickness of up to 1000 m (Ricketts, 1986). On the basis of preserved flora, the formation has been dated as Middle Eocene (Ricketts and McIntyre, 1986). The fossil forest site lies within the Upper Coal Member of the formation (Ricketts, *this volume*) and is represented by several facies associations.

The area has an arctic climate, characterized by long, cold winters and short, cool summers. Eureka, the nearest weather station, has a mean annual temperature of -19.7°C and annual precipitation of 64 mm, 64 per cent of which falls as snow. The warmest month of the year is July, with a mean daily temperature of 5.4°C; the coldest month is February, with a mean daily temperature of -38.0°C (Atmospheric Environment Service, 1982).

The study area is underlain by continuous permafrost. Active layer thickness ranges from 20 to 80 cm, depending on the moisture regime and the texture of the material. Contemporary soils in the area are Cryosols (Agriculture Canada Expert Committee on Soil Survey, 1987). A soil moisture deficit is produced through evaporation, so that many contemporary soils exhibit surface crusts of salt crystals.

MATERIALS AND METHODS

Method of sampling and location of sample sites

White layers occur throughout the exposed sections of the Buchanan Lake Formation. These layers are up to 50 cm

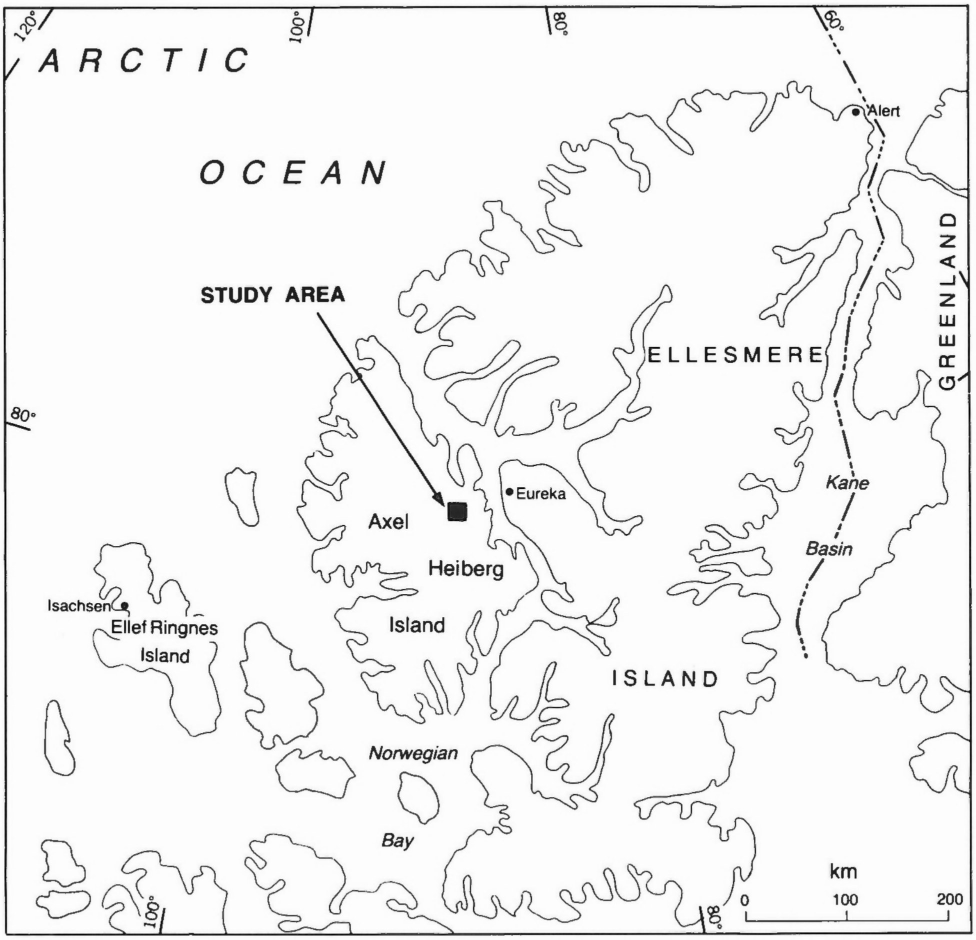

Figure 1*. Location of the fossil forest study site in the Geodetic Hills area, Axel Heiberg Island.*

thick, but are dominantly 5 to 20 cm thick. The white layers are found in several situations: between loamy mineral material; with organic material above and gray, reduced, silty mineral material below; or sandwiched between two distinct organic layers. None of the white layers examined was underlain by a Podzolic B horizon. Coarse mottles occurred in some of the thicker white layers, suggesting that the material had been exposed to wet conditions. The frequency of these layers was examined in a randomly selected 22 m vertical section. Seven white layers ranging from 5 to 20 cm thick were encountered in this section (Tarnocai and Smith, *this volume*, Fig. 6).

Thirteen white layer samples (W1-W13) were collected from exposed portions of two transects (A and B, Figs. 2 and 3); they are described in Table 1. Another white layer sample (W14) was collected from a nearby pedon, 12-88-71 (Detail No. 1, Fig. 6, Tarnocai and Smith, *this volume*). Before sampling, the white layer portion of the section was cleaned and dug back 10 to 20 cm in order to avoid contamination by slumped material. All samples were stored in plastic bags before being shipped to the laboratory for analysis.

Analytical methods

Eight samples selected from those collected were subjected to particle size distribution analysis using a Malvern Master Particle Sizer M3.1 courtesy of Dr. Luba S. Kotlyar, Division of Chemistry, National Research Council of Canada.

After each of the original samples was homogenized, 1 to 2 g were taken as a representative sample and sized down to <45 µm by a mechanical grinder. This rendered it suitable for X-ray powder diffraction (XRD) analysis, determination of loss on ignition (LOI) and chemical dissolution tests.

For XRD analysis, a Scintag PAD V automated powder diffractometer equipped with a graphite monochromator using cobalt radiation was employed on all 14 ground samples. A sample weighing 150 to 200 mg was mounted in a Plexiglas sample holder and scanned through diffraction angles ranging from 2° to 82°. LOI was determined by the difference in weight before and after heating a sample in a platinum crucible at 1000°C for 1 h.

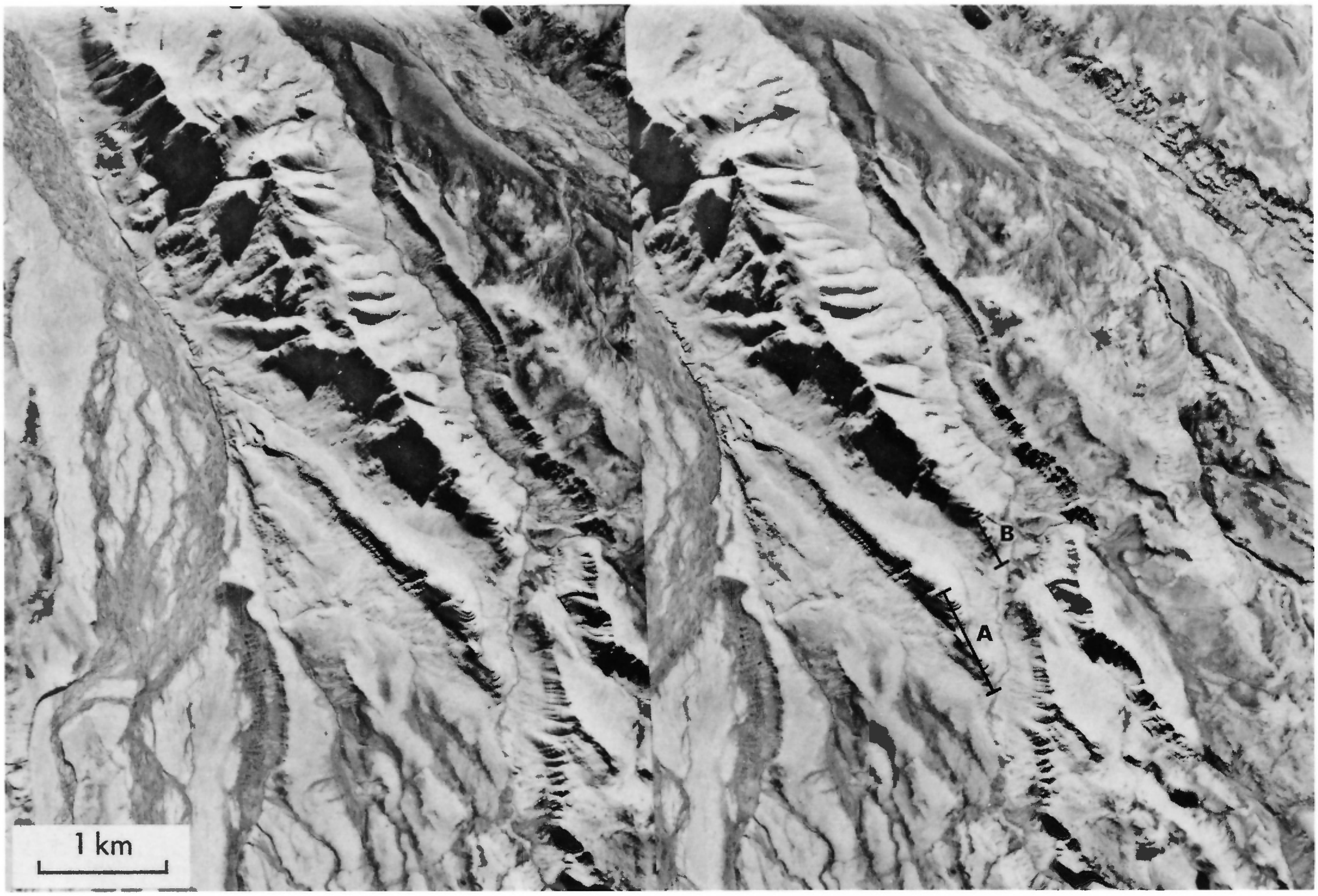

***Figure 2**. A stereopair of the fossil forest study area showing the locations of transects A and B.*

From XRD analysis and LOI determination studies of the 14 samples, it was found that quartz was the major mineral component. Samples W3, W4, W5 and W13 had the greatest amount of quartz. These four samples were then subjected to further analysis, including a chemical dissolution test with a hot 0.5M NaOH solution. This was applied to these selected samples to determine if noncrystalline silica was present.

A preliminary examination of thin sections of samples W3, W4, W5 and W13 with an optical microscope showed that most quartz crystals exhibited an uncommon morphology. The basal plane was better developed than the prismatic and pyramidal planes with concoidal fracture, as is normally observed in quartz. To our knowledge, substantial amounts of quartz with this morphology have never been reported.

It is possible that the platy crystal habit was inherited from tridymite when tridymite was transformed into quartz. Such a case may be seen in volcanic rock, but on a small scale. Therefore, it is unlikely that a large quantity of platy quartz crystals was transformed from tridymite. To define this peculiar morphology, unground original material of the four samples was examined under a scanning electron microscope using an Akashi 1S1 - DS130 instrument. A very dilute suspension was prepared using a Vortex mixer and an ultrasonic cleaner. A few small drops were placed on an aluminum stud and dried under ambient conditions. Further support for the specific crystal habit of the quartz studied may be obtained by comparing the peak intensity of its 003 reflection, I(003) at 1.80Å, with the peak intensity of its 112 reflection, I(112), at 1.82Å. X-ray diffraction patterns of unground samples W3 and W5 were recorded from 58° to 60° (2Θ, CoKα) with a slow scanning speed, and their I(003)/I(112) ratios were evaluated. These ratios were compared with that of a Brazilian quartz, which was used as a standard.

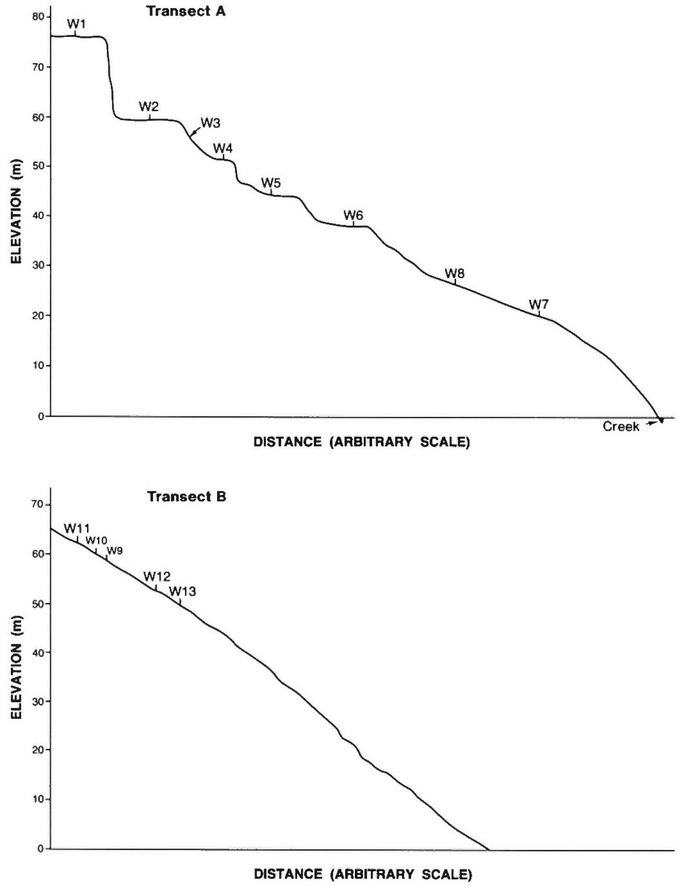

__Figure 3__. Cross-sections of transects A and B, showing the locations of the sample sites.

A subsample from each of the 14 white layer samples was suspended in distilled water and placed on a microscope slide to dry in air. The subsamples were examined using a polarizing light microscope (Ortholux II Pol-BK) at magnifications ranging from x25 to x630.

A subsample from layer W3 was air dried and impregnated with polyester resin, and a thin section was prepared. The constituents were examined for shape, size relationship and internal morphology. In addition, comparisons were made with scanning electron micrographs.

RESULTS AND DISCUSSION

Particle size distribution

Figure 4 shows particle size distribution curves of five white layer samples that were obtained from transect A. These curves show a polymodal distribution, not a normal distribution, suggesting complex disposal conditions and some infiltration of fines from different sources. Three distinct maxima were observed in all five samples and classified by ranges; 6 to 12 μm, 25 to 36 μm and 75 to 115 μm. Cumulative particle size distribution curves given in Figure 5 indicate that parti-

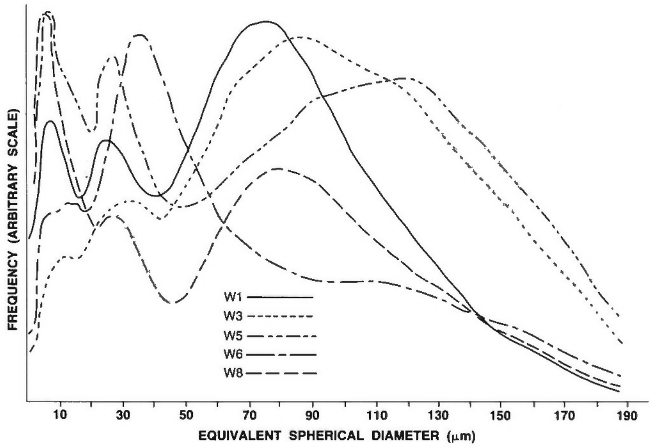

__Figure 4__. Particle size distribution (frequency) curves of samples W1, W3, W5, W6 and W8 from transect A.

TABLE 1

Description of white layer samples, fossil forest area, Geodetic Hills, Axel Heiberg Island

Sample No.	Transect	Elevation* (m)	Munsell colour (dry)	Comments	Remarks
W1	A	78	10YR 7/2	Top of Fossil Forest hill	-
W2	A	63	10YR 8/2	White layer plateau	-
W3	A	57	7.5YR 8/0	Jack McMillan's site (layer D)	+
W4	A	52	7.5YR 8/0		+
W5	A	45	7.5YR 8/0		+
W6	A	39	10YR 7/1	Light-coloured layer under peat	+ †
W7	A	21	10YR 7/2	Discontinuous layer	+
W8	A	27	10YR 7/1	Clayey layer with white coatings	+ †
W9	B	59	7.5YR 8/0		
W10	B	61	10YR 7/2		
W11	B	63	10YR 7/2		
W12	B	53	10YR 7/1	Some white material in a dominantly mineral deposit	† x
W13	B	50	10YR 7/1		+
W14	–	85	7.5YR 8/0	White layer from pedon 12-88-71	

* relative elevation above camp (measured by altimeter); + under peat layer; x overlain by mineral layer; - exposed on the horizontal surface (overlying material eroded); † not a typical white layer (mixed with other mineral material)

cles smaller than 50 µm make up more than 70 per cent of all samples, except sample W3, in which they make up about 55 per cent. Although it seems that there is a considerable variation among the polynomial distributions in Figure 4, a statistical test did not indicate the presence of significant variation since their bimodality indices (Sahu, 1964) ranged from 1.78 to 1.98. If sample W3 is excluded, the indices occur in the even narrower range from 1.93 to 1.98. Consequently, in this aspect, sample W3 should be considered unique among the samples along transect A.

Particle size distribution curves of three samples that were obtained from transect B (samples W10, W12 and W13) are given in Figures 6 and 7. Like transect A samples, they did not show normal distribution. All three of these samples had two distinct maxima with an obscure third maximum or hump. If they are treated as having bimodal distribution, by ignoring the third hump, the distributions of samples W10 and W13 are alike, with their two maxima appearing in the 24 to 27 µm and 73 to 76 µm ranges. The two maxima in sample W12, however, are at much smaller particle sizes, i.e., at 5.9

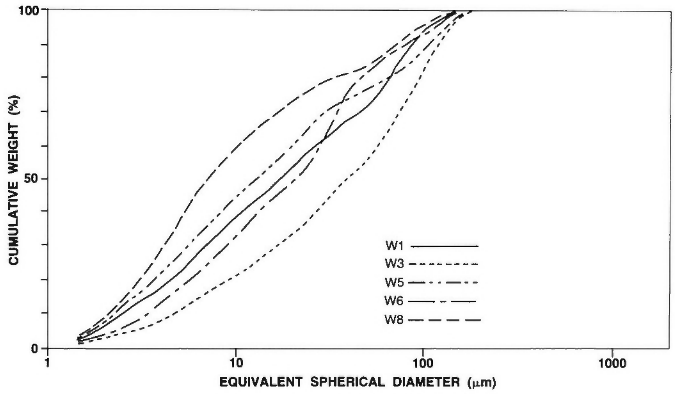

Figure 5. Cumulative percentage curves of particle sizes in samples W1, W3, W5, W6 and W8 from transect A.

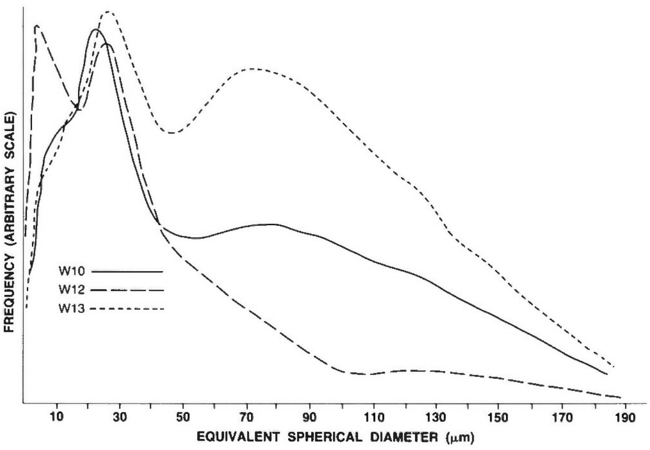

Figure 6. Particle size distribution (frequency) curves of samples W10, W12 and W13 from transect B.

μm and 27 μm. Cumulative particle size distribution curves given in Figure 6 indicate that particles smaller than 50 μm make up more than 70 per cent of those samples and more than 90 per cent of sample W12. A statistical test gave bimodality indices of 1.66, 2.19 and 1.44 for samples W10, W12 and W13, respectively, indicating that some significant variation existed among their distributions. The polymodal distribution suggested by these data indicates that the mineral components making up the white layers were derived from various sources and deposited under various conditions. A further investigation is being planned to determine concentrations of mineral components in each of the major particle size ranges and to characterize the quartz in each range.

Mineral composition

The XRD analysis showed that highly crystalline quartz was a major mineral component in all 14 samples. Kaolinite was also present in all samples, but its quantity varied from trace to moderate. The amount of quartz in the samples was estimated from the intensity of the quartz reflection at 4.26Å by comparing it with that for a quartz standard (100%). The estimated amounts are given in Table 2. The quartz content ranged from 56 per cent to 98 per cent (see Fig. 8 for the latter example). The other mineral components detected were mica, 14-Å minerals such as chlorite and vermiculite, interstratified clay minerals, feldspar, cristobalite, zircon, rutile and anatase. These minerals were present in trace to minor quantities.

There was a trend for the LOI (data given in Table 2) to increase with decreasing quartz content. Although samples with lower quartz content had darker colours, the LOI could not always be correlated with the darkness (colour value) of the samples, possibly because of the effect of organic matter present. This is because ignition during the LOI determination causes the combustion of organic matter and the dehydration and dehydroxylation of hydrous phyllosilicates such as most of the clay minerals.

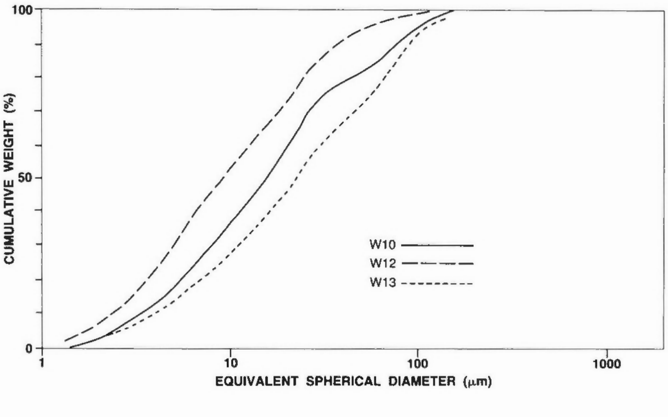

Figure 7. Cumulative percentage curves of particle sizes in samples W10, W12 and W13 from transect B.

Four of the 14 samples tested (W3, W4, W5 and W13) contained more than 90 per cent quartz. The crystallinity of the quartz in these four samples was examined chemically. The chemical dissolution treatment with NaOH indicated the presence of some soluble (noncrystalline) silica in the samples. The amount ranged between 2.1 and 3.1 per cent occurring as SiO_2, suggesting that a small amount of noncrystalline silica was associated with highly crystalline quartz. In a Brazilian quartz used as standard, the amount of soluble silica was less than 0.2 per cent.

Crystal morphology of the quartz

Scanning electron microscope observations (Fig. 9) were focused mainly on crystal particles smaller than 10 μm. It is evident from Figure 9 (especially Fig. 9A, B, D, E, F, and G) that a considerable number of clay-size particles were present in association with the fine and medium silt-size particles, which formed one of the major particle size (distribution) groups. The clay-size particles were attached to the silt-size particles (Fig. 9A, B, D, F, and G) or, less frequently, scattered more freely in the space between aggregated particles. One of the most striking features was the significant number of quartz crystals occurring as platy and flaky particles without well defined outlines (Fig. 9B, C, G, and H). This feature, and the fact that many particles placed on a glass slide behaved like optically isotropic substances in polarized light, indicate that the dominantly developed face of the particles was the basal

TABLE 2

Analysis of the white layer samples

Sample	Intensity of quartz reflection at 4.26Å (CPS)	Estimated amount of quartz (%)	Loss on ignition (%)	Minor mineral constituents
W1	162	65	4.18	F, M, K
W2	151	61	5.13	K, M, 14-Å , IN, F
W3	243	98	1.24	K
W4	228	92	2.23	K, M
W5	238	96	1.29	K
W6	166	67	5.94	K, 14-Å , M, Z, A
W7	168	68	5.73	K, M, R, 14-Å
W8	195	79	3.88	K, M, A, R
W9	179	72	4.21	K, M, IN, A
W10	194	78	3.58	K, M, IN, F, A
W11	160	65	4.84	K, M, IN, A, Z
W12	140	56	6.43	K, M, IN, Z, A, R
W13	228	92	3.11	K, IN, M, Cr
W14	196	79	3.16	K, IN, A, M, R, Cr

K - kaolinite; F - feldspar; R - rutile; M - mica; Z - zircon; Cr - cristobalite;
IN - interstratified clay minerals; A - anatase; 14-Å - chlorite and/or vermiculite.

plane normal to the c-axis of the quartz. This crystal habit is not common in quartz.

The peculiar crystal habit in these samples should contribute to enhancement of the intensity of the XRD reflection responsible for the dominant basal plane. The 003 reflection at 1.80 Å was the only observable basal reflection for quartz. Partial XRD patterns of a quartz standard and unground sample W3 are given in Figure 10A and B, respectively. As seen in Figure 10A, the 003 reflection is usually very weak and the intensity ratio, I(003)/I(112) is 0.11, whereas the ratio for W3 is 0.53 and the ratio for W5 is 0.48. The ratio of the quartz in these samples is approximately five times larger than the standard, showing that the basal plane of many quartz particles in the samples was abnormally developed with respect to other crystal planes.

Micromorphological attributes of the white layer

The dominant characteristic of the white layers is the crystalline, plate-like morphology of the quartz particles. The particles are predominantly less than 30 μm in size (Fig. 11B), often subangular to subrounded with a distinct platy appearance (Figs. 9E, G; 11B, C). Of special note are the occurrences of unaltered crystal faces shown in Figure 9B, E and suggested in Figure 11A, D.

The quartz particles, especially the platy ones, are extremely transparent. When the samples are examined with a light microscope it is found that inclusions and impurities are extremely rare to nonexistent. Smaller crystals are usually visible beneath the larger crystalline particles.

Occasional crystals in Figure 11C, D show evidence of surface variation; this morphology was related to the presence of extremely fine particles adhering to the surface, as is shown especially well in Figure 9F. These variations also result from the adherence to the surface of plate-like features, as shown in Figure 9B, C, and E. This morphology may have resulted from weathering processes.

The quartz particles >100 μm in Figure 11A tend to be subangular to angular and show occasional internal cracking

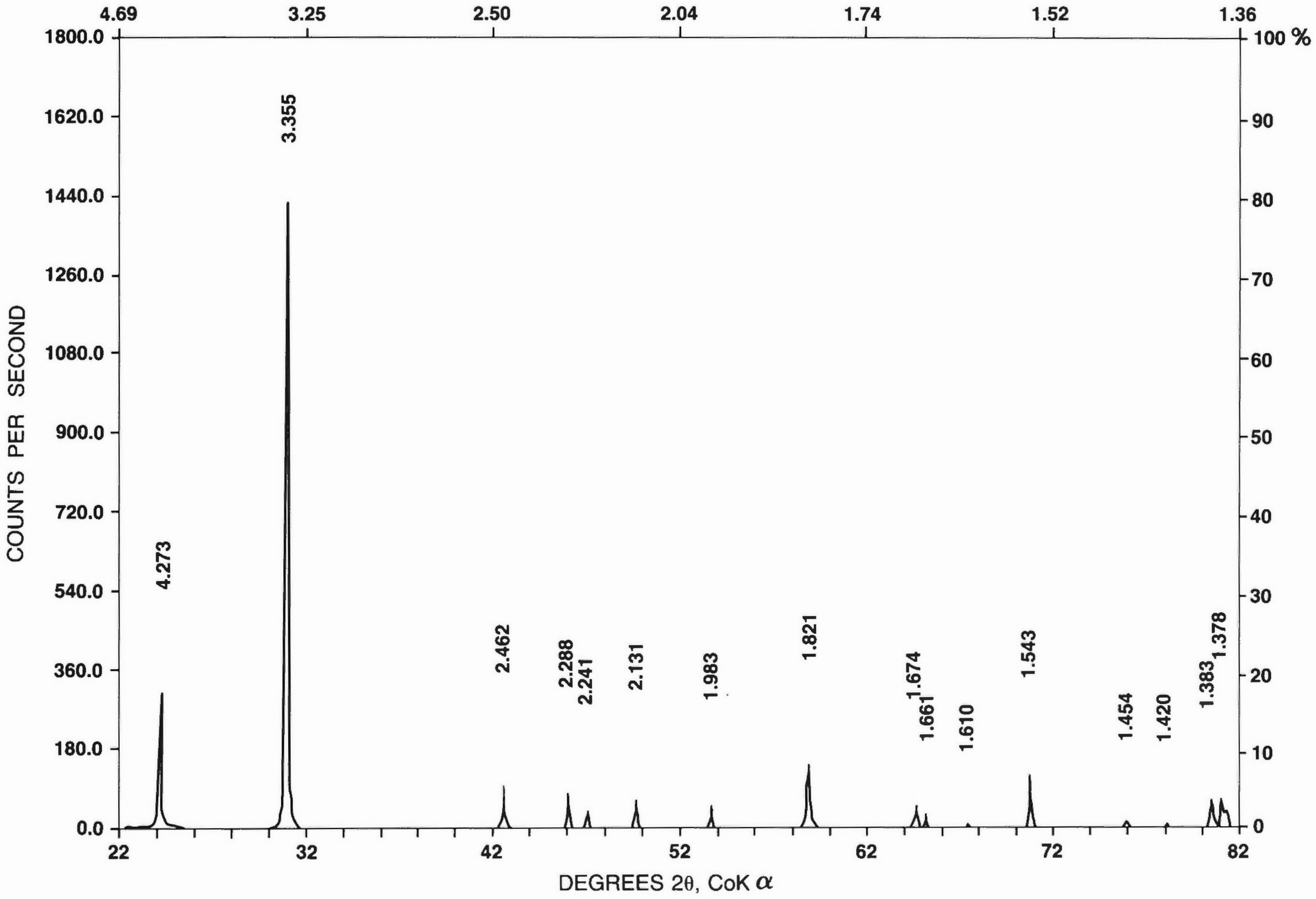

Figure 8. X-ray powder diffraction pattern of sample W3.

patterns, possibly the result of weakening by tectonic processes combined with later cryogenic processes. Such cracks cause breakup of the particles, resulting in angular particles ranging in size from 40 to 100 µm. These particles tend not to be plate-like in appearance unlike those <50 µm, suggesting an alternative genetic process. Single grain particles of quartz dominate, but there are also rare occurrences of chert and quartzitic lithofragments, suggesting inclusion of some material from quartz-rich source areas.

Included among the plate-like crystals are rod-like fragments (Fig. 11B, C, D), 5 to 12 µm in length. At the bottom right of Figure 9, D is a similar rod-like structure. The origin of these fragments is unknown — it may be biological, such as diatoms or phytoliths, or precipitation of soluble salts.

DISCUSSION

The particle size distribution (Fig. 4) of the white layer samples does not show a characteristic Normal Gaussian distribution (Griffiths, 1967), suggesting that complex depositional events have taken place. In all 14 samples studied, most of the quartz crystals had a platy or flaky morphology. This type of crystalline habit has a large I(003)/I(112) ratio, the ratio of the basal reflection of quartz to the pyramidal reflection. The observed ratio of 0.5 is much larger than the ratio of 0.1 normally found in quartz crystals. The purity of the material, together with the plate-like morphology of the quartz crystals and the dominant subangular shape of the plates, suggests that the white layer material may not be the direct result of fluvial or eolian processes. These processes tend to produce quartz grains with rounded edges, not plate-like particles like those in the material studied.

In order to determine whether the white layer material has a pedological origin, as suggested by Francis and McMillan (1987), it is first necessary to consider both the soil genesis and the soil horizon sequences associated with eluviated soil horizons. Leached horizons are commonly found in contemporary Podzols, Brunisols and Luvisols (Agriculture Canada

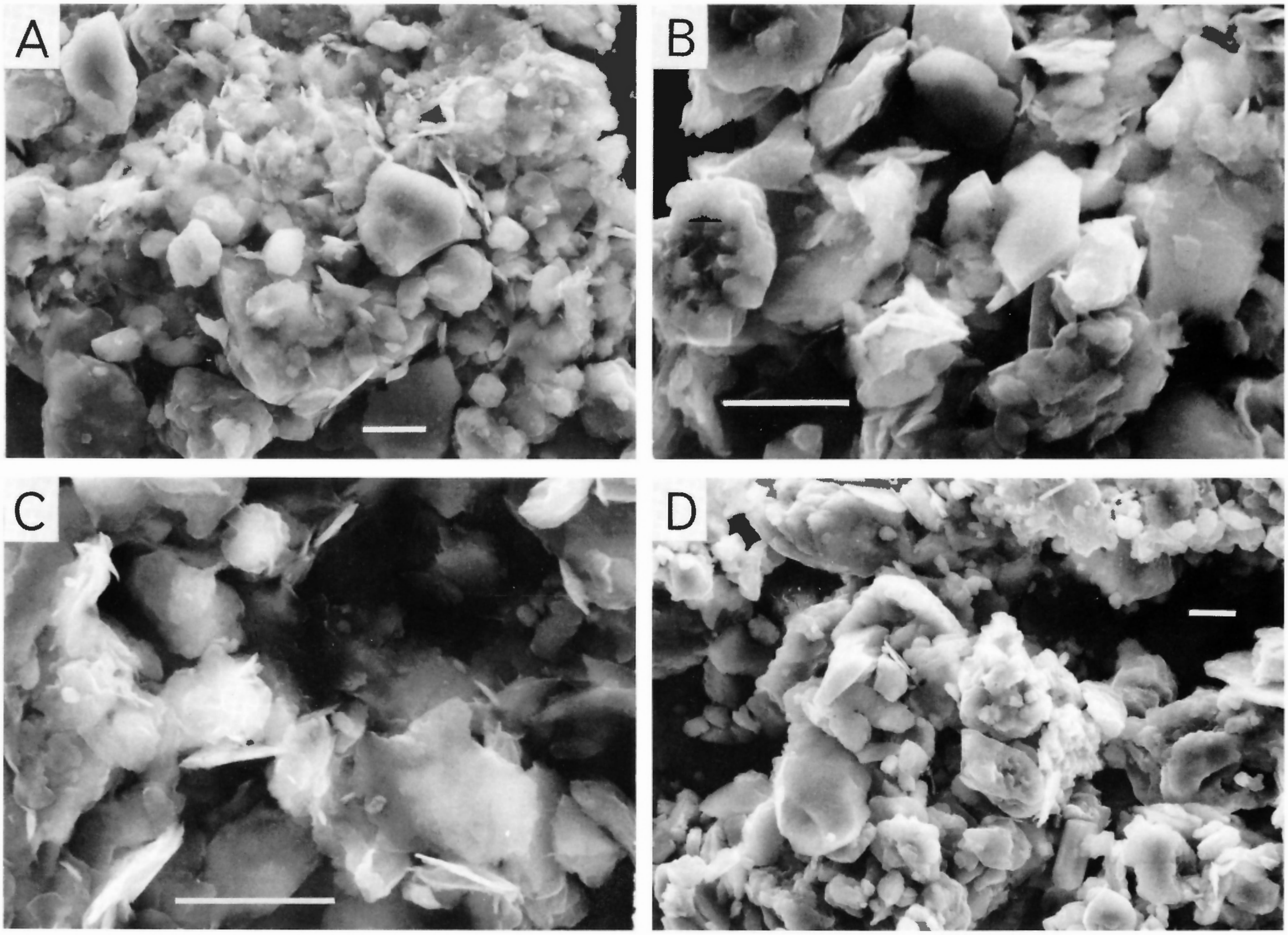

Figure 9. Scanning electron micrographs of samples W3 (A, B and C) and W5 (D). Scale bar = 5 µm.

Expert Committee on Soil Survey, 1987). The iron and aluminum are removed from the leached (Ae) horizon by organic acids produced by decomposition and accumulate in the Bf and Bm horizons in the form of organo-mineral complexes. The clay and other leachable mineral and organic materials are removed from the leached horizon of the Luvisols and accumulate in the Bt horizon. As a result of these pedological processes, such soils have a generally whitish coloured, leached (Ae) soil horizon underlain by a brownish, accumulative Bf, Bm or Bt horizon. No such pedological sequence was found under well drained conditions during the field examination of the white layers (Tarnocai and Smith, *this volume*). The white layers were underlain by either strongly gleyed and mottled silty mineral horizons (Cg) or peaty organic horizons (O). This suggests that the white layers developed in a wet or water-saturated environment and were not associated with accumulative soil horizons. These data and the crystal morphology of the quartz in the white layers indicate that a pedological origin is unlikely.

The possibility of a biological origin for the white layer material was given careful examination, especially because of the peculiar crystal habit of the quartz. Biological silica (phytoliths) originates from a variety of plant species. Silica constitutes the skeletal structure of diatoms, which take many, very distinct, forms characteristic of particular diatom species. Plant opal, another common type of biological silica, accumulates in the cell structure of plants (e.g., trees, shrubs and grasses). Plants take up monomeric silicic acid along with other nutrients from soil solutions and transport them through vascular cells to the leaf tissues. Silica is then deposited in the leaf tissue as plant opal. As a consequence of the accumulation of silica in the plant cells, the morphology of the plant opal is characteristic of the plant species.

Biological opal, which is produced by both diatoms and higher plants, is composed of amorphous silica. Wilding and Drees (1974) found that leaf tissues contain both amorphous and crystalline silica. The crystalline phase of this silica

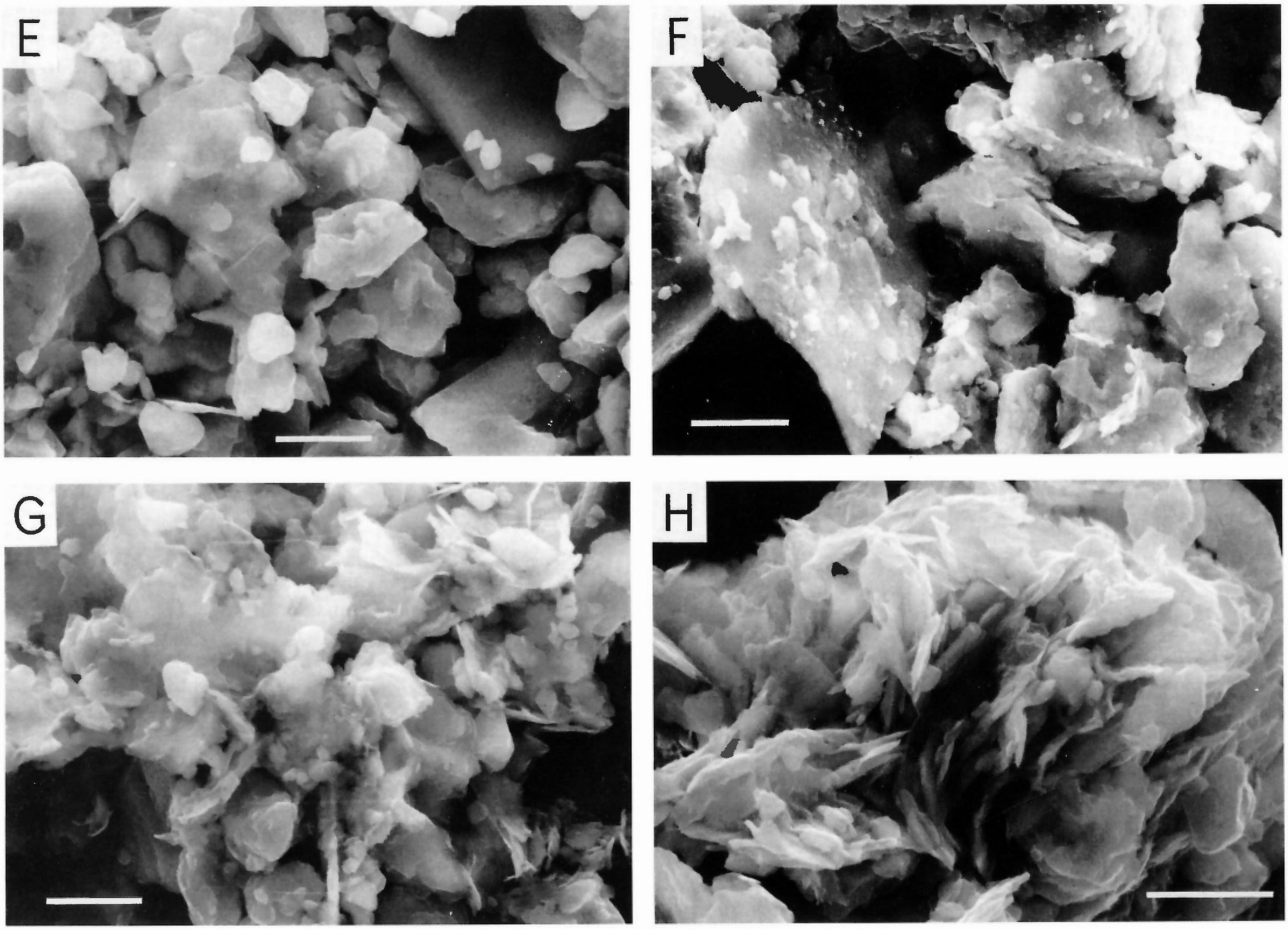

***Figure 9 (cont'd).** Scanning electron micrographs of samples W5 (E), W4 (F), W13 (G and H). Scale bar = 5 μm.*

originates from trees and appears as thin, coarse-textured, tabular or slightly scrolled silica sheets.

Baker (1960) reported opal phytoliths from sediments of Paleocene age in western Victoria, Australia. These phytoliths were generally rod-like, ranging from 0.025 mm long and 0.006 mm wide to 0.030 mm long and 0.015 mm wide.

Baker (1960) also states that opal phytoliths do not persist for more than approximately 1000 years under conditions existing in present-day soils in Victoria. Depending on the chemistry of the soil, the opal phytoliths are generally taken into solution. Baker (1960) did suggest that: "Under geological conditions, however, where they can sometimes be sealed relatively rapidly and thus protected from attack by the more severe corrosive agents, some examples can be preserved for several millions of years."

Baker (1960) also states that amorphous opal can go through a dehydration process and then be converted into crystalline quartz. But, as he stated in his paper, "Given the right conditions for dehydration, such a conversion might be expected among the fossil opal phytoliths observed in the older Tertiary sediments. If such a process has occurred, the newly formed products have not retained any shapes that would identify them with the original forms of opal phytoliths."

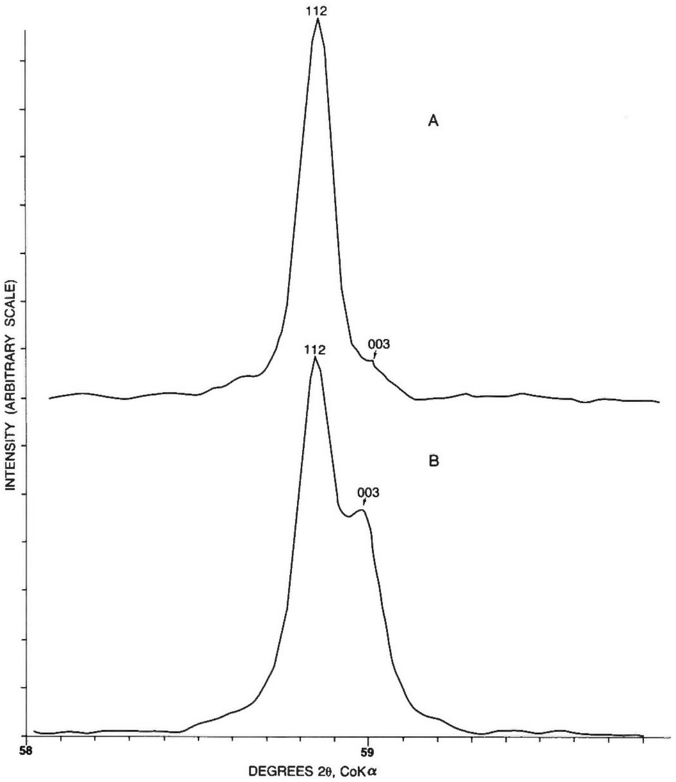

***Figure 10**. Partial x-ray powder diffraction pattern comparing the peak intensities of the 112 and 003 reflections of a Brazilian quartz standard (A) with those of sample W3.*

Circulating pore water reacts with the buried opal phytoliths, causing their transformation to crystalline quartz (Blatt et al., 1972). Opal phytoliths contain 2 to 13 per cent water and, even in the absence of pore water, there is enough water present within the opal to start the opal-quartz transformation (Blatt et al., 1972). Quartz is anhydrous and thus a certain amount of energy is required to expel the water from the opal during crystallization. Ernst and Calvert (1969) determined the rate constant for the opal-quartz transformation in neutral solution with no circulation of water. They suggested that at 200°C complete conversion of opal phytoliths to crystalline quartz takes 47 a; at 100°C, 36 ka; at 50°C, 4.3 Ma; and at 20°C, 180 Ma.

The high quartz content of the white layer material and the peculiar crystal habit of the quartz in this material suggest that it may have originated from opal phytoliths produced by the Eocene vegetation growing in the area. As a result of continual fluvial deposition, the opal phytoliths beds possibly were quickly sealed and corrosive processes were unable to dissolve the opal phytoliths. It is then possible that after burial the opal phytoliths deposit was converted to quartz by the dehydration process. Temperature conditions were favourable within the limits suggested by Ernst and Calvert (1969) since lignite, which also occurs in these deposits, requires temperatures of 25° to 100°C to form. From the degree of lignification it was estimated that the fossil forest deposits were exposed to temperatures of approximately 50°C (Ricketts, pers. comm., 1989). Some of the rod-like particles found in the white layer samples are probably derived from crystalline phytoliths produced by plants. These remain in their original form. The amorphous opal phytoliths, the dominant material in these layers, have lost all the original morphology, however, except platiness, during the conversion to crystalline quartz.

SUMMARY

1. Particle size distribution of the white layer material does not show Normal Gaussian distribution, suggesting a complex or unusual origin.

2. All white layer samples contain high amounts of quartz which has a platy morphology [high I(003)/I(112) ratio], unlike the morphology of commonly occurring quartz grains.

3. The particle size distribution and the shape of the quartz crystals suggest that the white layer material was not deposited by fluvial or eolian processes.

4. The lack of B horizons underlying the white layers and the platy morphology of the quartz suggest that the white

layers did not originate from pedological processes resulting from leaching.

5. The crystal habit of the quartz, the unusual particle size distribution, and the rod-like features found in the white layers suggest that the material originated from opal phytoliths. After burial, the amorphous opal phytolithic material went through a dehydration process that converted it to crystalline quartz. As a result of this conversion, nearly all phytolithic morphology was lost.

ACKNOWLEDGMENTS

The authors are grateful to the Polar Continental Shelf Project for providing the logistical support necessary for carrying out the field work. Special thanks are also due to M. Jaakkimainen of the Land Resource Research Centre (LRRC), Ottawa, for carrying out the chemical extraction and for the SEM and LOI analyses, to G.C. Scott of LRRC for carrying out the XRD analysis, to R. Guertin of LRRC for preparing the thin sections, and to Dr. L.S. Kotlyar of the National Research Council, Ottawa, for the particle size analysis.

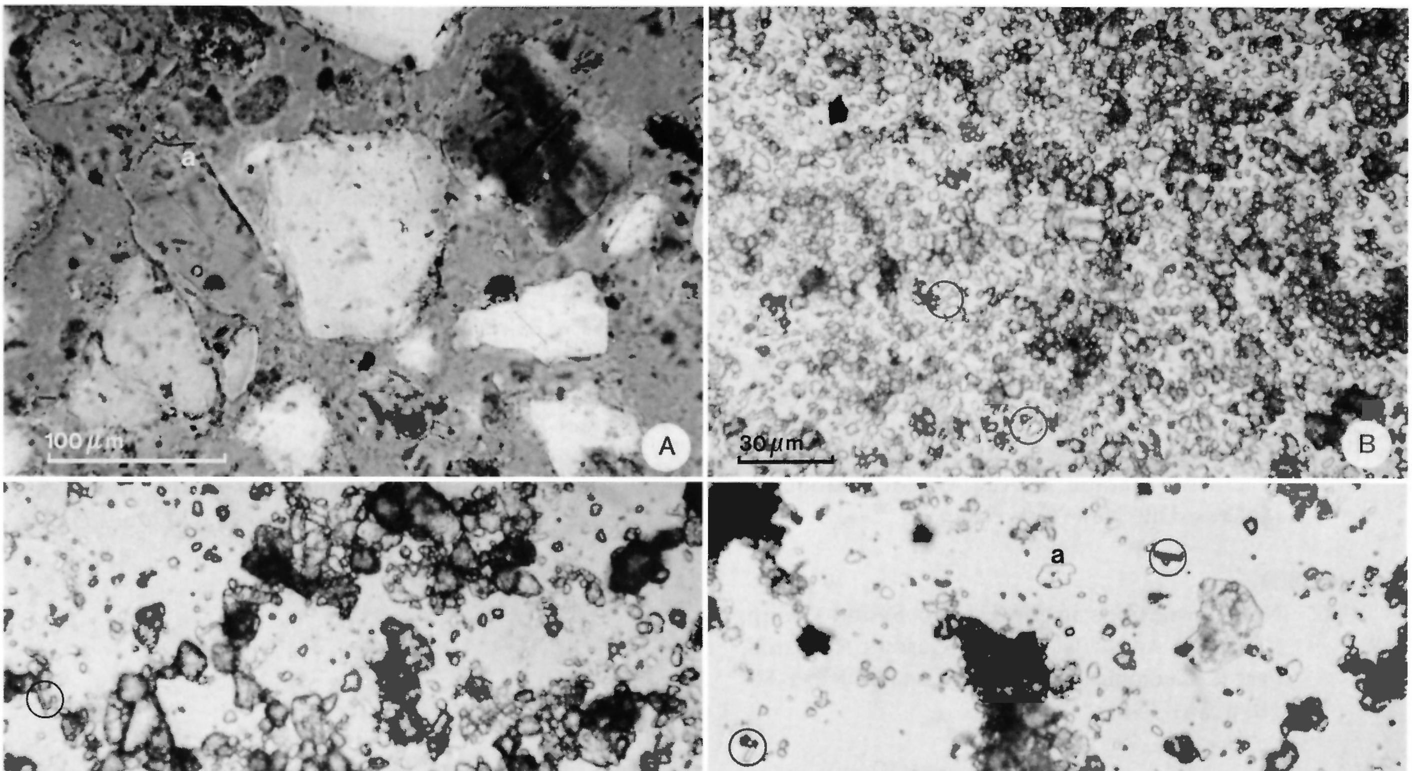

***Figure 11**. Morphological characteristics of the white layer particles: A, angular to subangular faces are dominant (indicated by "a"), occasional cracks are visible in interior of particles; thin section of sample W3, partially crossed polarized light; B, overview of slurry sample (W2) showing predominance of particles less than 30 μm (the circles indicate features of possible opal phytolith origin); partially crossed polarized light; C, additional view of slurry sample (W2) showing transparency and plate-like feature of particles; partially polarized light; D, view of slurry sample (W12) showing larger transparent particles with surface adhesion of smaller particles, rod-shaped particles (within circles), and distinct crystal faces (a) on occasional particles; partially polarized light.*

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AN INCIPIENT TERTIARY SOIL PROFILE FORMED ON GREY CLAY AND BIOGENIC PHYTOLITE (WHITE LAYER), AXEL HEIBERG ISLAND

A.E. Foscolos[1] and N.J. McMillan[2]

Foscolos, A.E. and McMillan, N.J., 1991. An incipient Tertiary soil profile formed on grey clay and biogenic phytolite (white layer), Axel Heiberg Island. In Tertiary Fossil Forests of the Geodetic Hills, Axel Heiberg Island, Arctic Archipelago, R.L. Christie and N.J. McMillan (eds.); Geological Survey of Canada, Bulletin 403, p. 201-207.

Abstract

Three samples of a soil profile beneath 5 cm of leaf litter were chemically analyzed, and the clay minerals studied. The parent material is clay and biogenic phytolite. There is an increase in pH from the topmost layer to the bottom layer. Analyses indicate an increase in trace elements in the lower part of the profile. Smectite was present throughout the section but tended to be degraded in the centre of the profile.

The paleosol is weakly developed and is composed of an organic surface horizon underlain by a gleyed mineral horizon. This weak development was probably the result of insufficient time for stronger soil development. The grey colours were likely the result of wet ground conditions.

Résumé

On a soumis à une analyse chimique trois échantillons d'un profil de sol situé sous 5 cm de litière de feuilles, et étudié les minéraux argileux. Le matériau parental se compose d'argile et de phytolite biogénique. On a observé un accroissement du pH de la couche sommitale à la couche basale. Les analyses ont indiqué un accroissement des taux d'oligo-éléments dans la partie inférieure du profil. De la smectite était présente dans tout le profil stratigraphique, mais montrait une tendance à la dégradation au centre du profil.

Le paléosol est faiblement développé, et se compose d'un horizon superficiel organique recouvrant un horizon minéral gleyifié. Ce faible développement résulte probablement de ce que la durée d'évolution du sol a été trop courte. Les couleurs grises ont probablement été dues à la très forte humidité du sol.

INTRODUCTION

A Tertiary fossil forest site was discovered by geologists of the Geological Survey of Canada in the Geodetic Hills area of Axel Heiberg Island in the Canadian Arctic in 1985 (McMillan, 1986; Francis and McMillan, 1987). The site is characterized by remarkably well preserved stumps, logs, and forest floor litter in a bedded sandstone-siltstone sequence. Associated with the forest litter layers are paleosols and enigmatic thin "white layers" interpreted by Tarnocai et al. (*this volume*) as phytolites.

In July 1986, three samples of a paleosol were taken by N.J. McMillan and Jane Francis below level "D" (a forest litter horizon) at the Geodetic Hills fossil forest site (Francis and McMillan, 1987). Care was taken to cut the samples from deep in the hillside to avoid contamination and to obtain true fossil soils not overprinted by present-day soil-forming processes. The sampling locality is shown in Figure 1. Closer views of the sampling site are shown in Figure 2. Figure 3 is a drawing of the profile. An account of analytical results is presented here.

DESCRIPTION OF SOIL PROFILE

A slight visible gradational difference in the soil extends from an overlying woody forest floor litter (approximately 5 cm thick) to a depth of 60 cm. The top layer is 10 cm of friable, weakly banded, grey-brown clay with no mottling. It is in sharp contact below with a dominantly white quartz-rich layer 25 cm thick. This white layer is uniform from top to bottom and is interpreted by Tarnocai et al. (*this volume*) to be a biogenic phytolite. It is in sharp contact below with banded grey-brown clay, not unlike the material above, but slightly darker. The attraction of the site was the even white layer at a fairly constant 10 cm depth below the leaf litter.

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[1] Department of Mineral Resources Engineering, Technical University of Crete, 129 Eleftheriou Venizelou Street, 73133 Chania, Greece
[2] Geological Survey of Canada, Institute of Sedimentary and Petroleum Geology, 3303 - 33rd Street N.W., Calgary, Alberta T2A 2A7

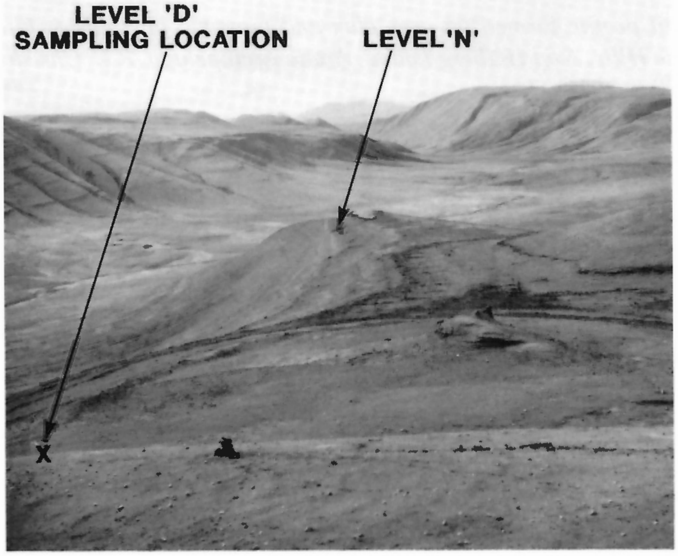

Figure 1. Geodetic Hill site, Axel Heiberg Island (see Francis and McMillan, 1987, for columnar section). The "D" level is 43 m above "N".

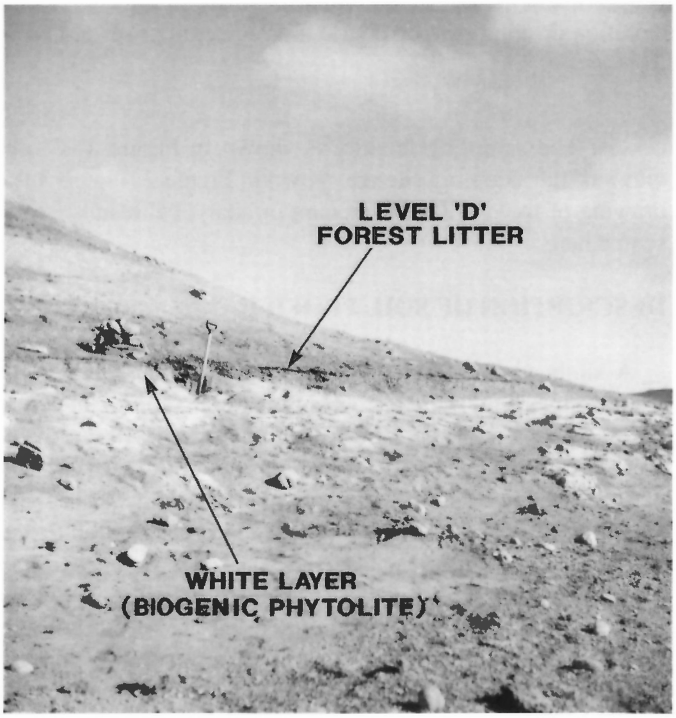

***Figure 2.** Closer view of white layer and forest litter, level "D" (see Francis and McMillan, 1987, for relationship of level "D" and level "N").*

No features were observed that would indicate the action of normal soil-forming processes. The "white layer" contains no evidence of leaching by way of colour change, mottling, sesquioxide blemishes or concretions. Further, it contains no soil structures, such as banded fabric. The layer immediately below seems to have received no contributions from above by leaching.

Three composite samples were taken at depths of 0 to 10 cm, 10 to 35, and 35 to 60 cm, respectively (see Fig. 3).

MINERALOGY AND CLAY MINERALOGY OF PALEOSOLIC HORIZONS

X-ray analysis of paleosolic horizons indicates that quartz is the most predominant mineral throughout (Table 1; Figs. 4 - 6). Lattice silicates and mixed layer clays increase with depth. The latter are present in the lowermost horizon.

In the <2 μm fraction, expandable silicates, mixed layer silicates, illite, kaolinite, and anatase are encountered throughout the profile (Figs. 4 - 6).

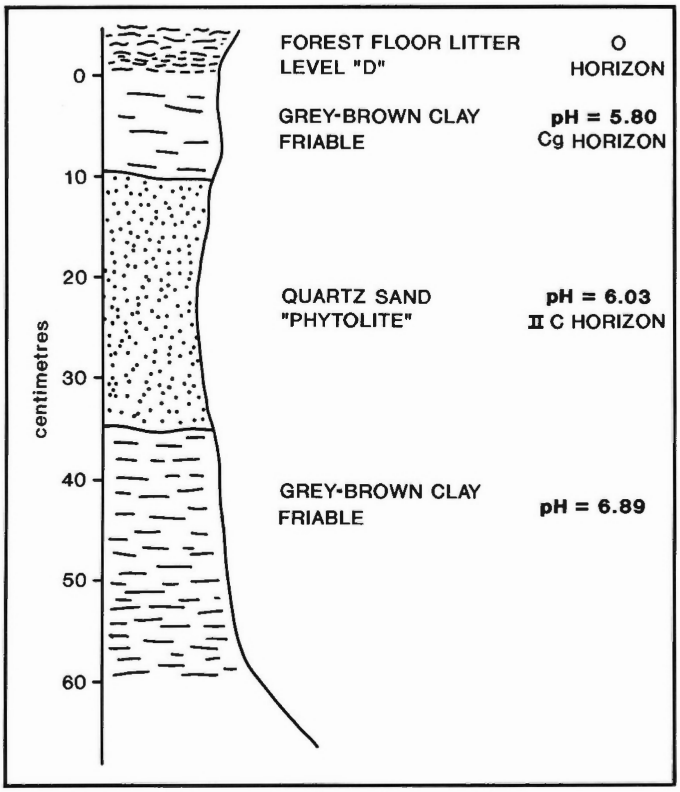

***Figure 3.** Soil column, level "D" [see Francis and McMillan, 1987, for location; see Tarnocai, et al. (**this volume**), for information on phytolite.] (Soil horizons suggested by C. Tarnocai, pers. comm.)*

ELEMENTAL ANALYSIS OF THE PALEOSOLIC HORIZONS AND THEIR <2 μm FRACTIONS

Measurements of hydrogen ion concentration were carried out on 1:5 soil extracts. The results show that pH values increase with depth (Table 2). From the top down, 0-10 cm is acidic, the next 25 cm slightly acidic, and at 60 cm the pH is neutral. It appears as though the ancient floor litter had imparted the pH values and that they were preserved in the paleosolic section.

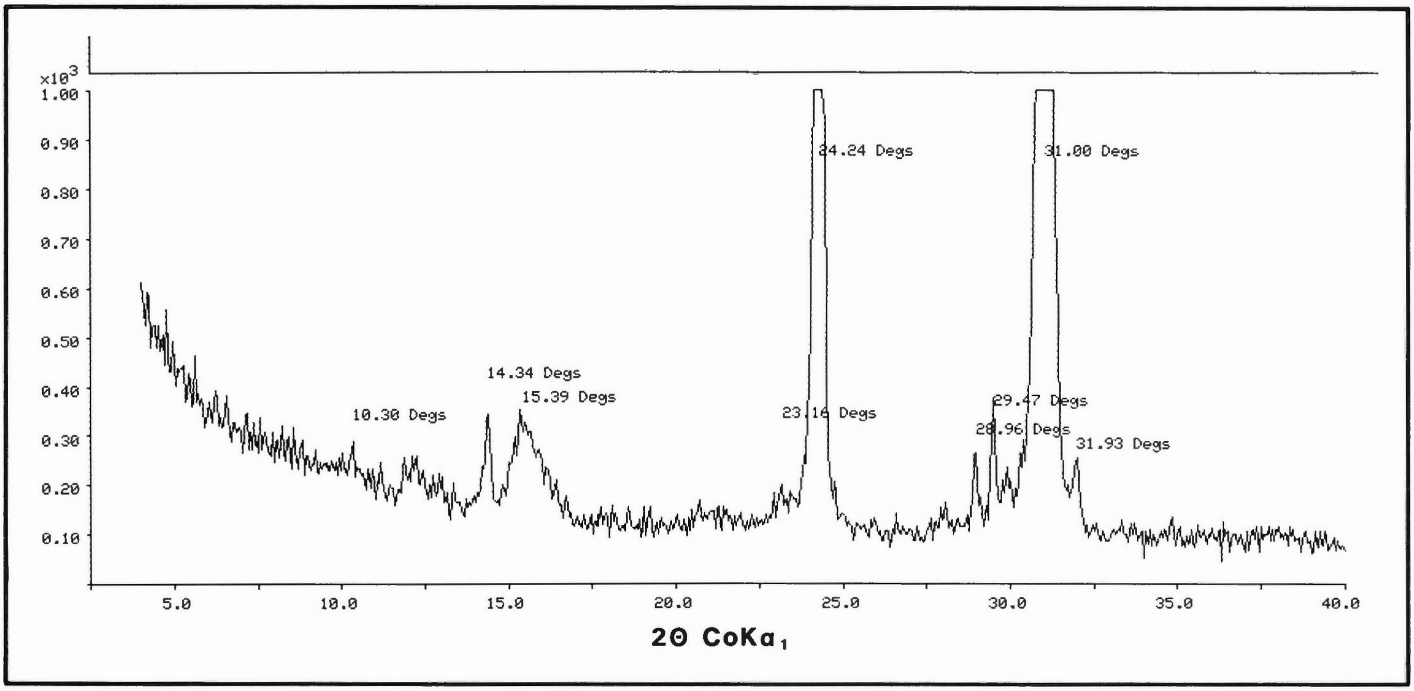

Figure 4. *X-ray diffractogram of soil from forest litter to 10 cm depth.*

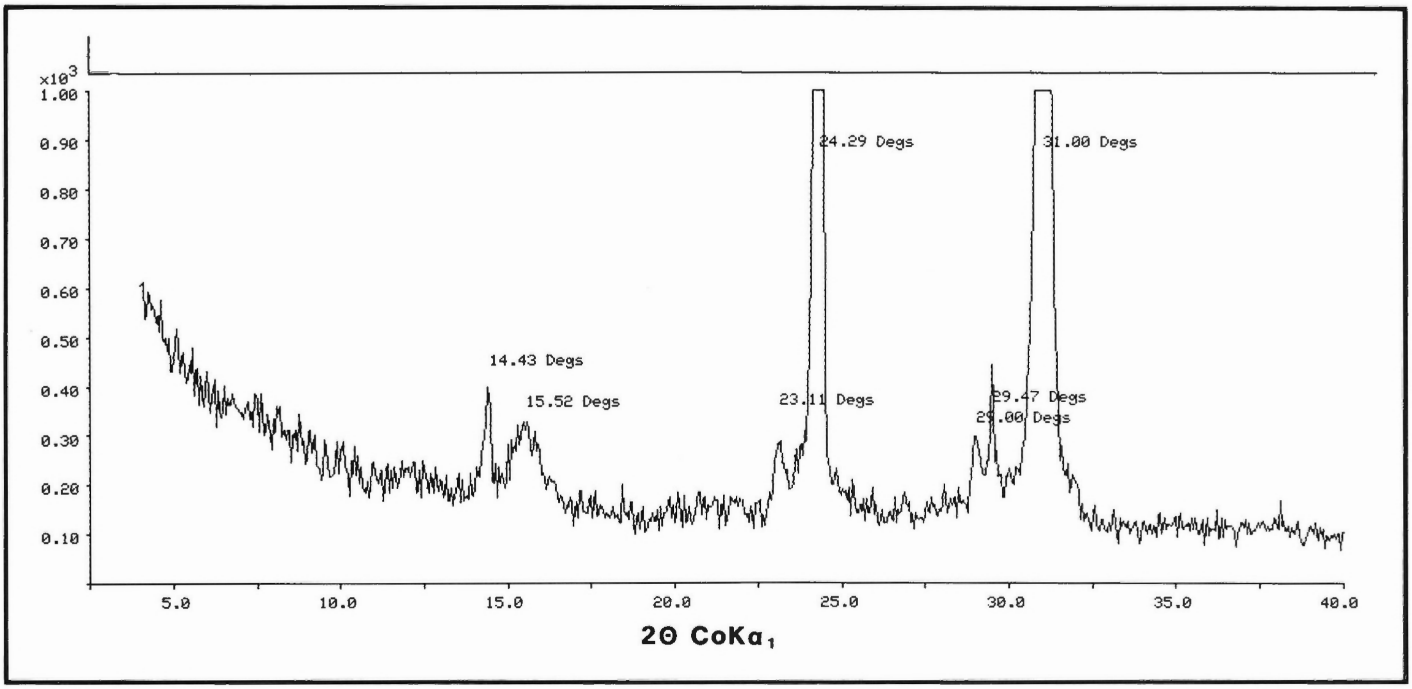

Figure 5. *X-ray diffractogram of soil from 10 to 35 cm depth interval (white layer).*

The elemental analysis of the paleosolic horizons yields high SiO_2 values, varying from 78.08 to 96.10 per cent (Table 2). This is attributed to the presence of quartz (Table 1). Trace element analysis on the same horizons reveals a simple increase with depth (Table 3). Elemental analysis of the Ca-saturated <2 μm fraction indicates that the uppermost and lowermost horizons have higher loss on ignition (L.O.I.) values than the middle layer (Table 4). This implies that hydrous layer silicates are abundant in the upper and lower horizons. Alternatively, the higher loss of ignition values for the grey clay could be due to the organic matter content of this horizon.

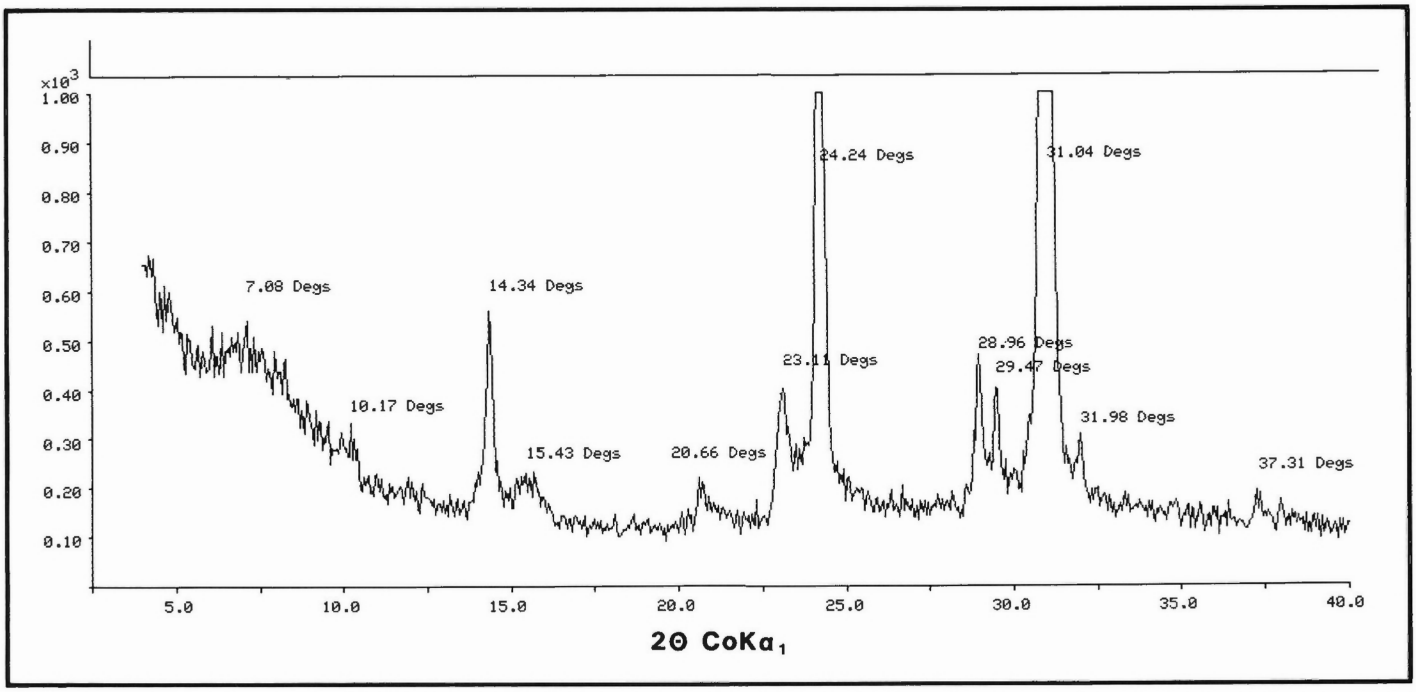

Figure 6. *X-ray diffractogram of soil from 35 to 60 cm depth interval.*

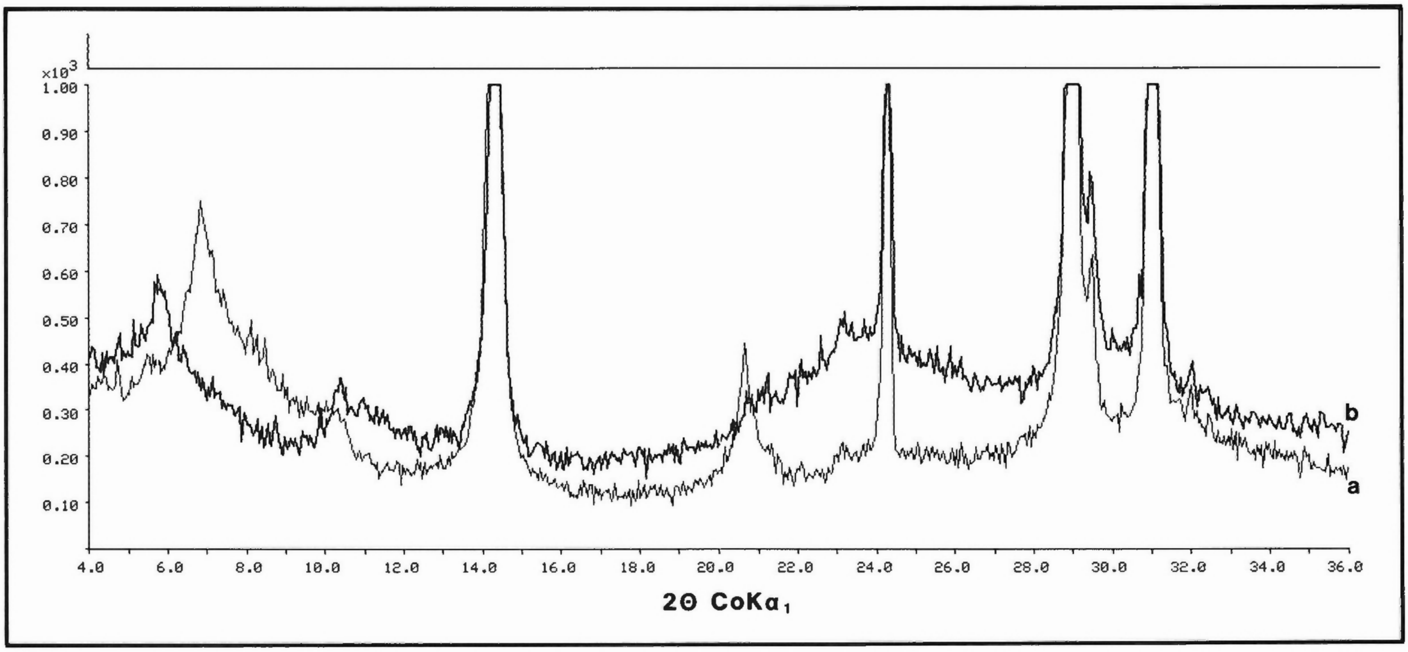

Figure 7. *X-ray diffractogram of the <2 μm fraction from forest litter to 10 cm depth. A, room conditions; B, glycerolation.*

TABLE 1

Semiquantitative mineralogical composition of paleosolic horizons as determined by X-ray diffraction

Lab. no.	Position of sample	ML and/or expand.	Illite	Kaolinite and/or chlorite	Quartz	Feldspar	Apatite and/or siderite
MCM10D	10 cm below coal bed D			3	97		
MCM35D	35 cm below coal bed D		1	2	96	1	
MCM60D	60 cm below coal bed D	5	2	6	85	1	1?

ML = mixed layered; expand. = expandable clays.

TABLE 2

Per cent elemental analysis of paleosolic horizons

Depth (cm)	SiO_2	Al_2O_3	Fe_2O_3	TiO_2	CaO	MgO	Na_2O	K_2O	P_2O_5	BaO	MnO	Loss on ignition	pH
0-10	88.20	3.25	0.66	1.63	0.09	0.13	0.00	0.28	0.03	0.03	<0.01	5.70	5.80
10-35	96.10	1.32	0.46	1.38	0.00	0.11	0.00	0.18	0.02	0.02	<0.01	0.40	6.03
35-60	78.08	7.38	5.80	1.55	0.22	0.38	0.07	0.74	0.06	0.03	0.07	5.60	6.89

TABLE 3

Trace elements of paleosolic horizons (ppm)

Depth (cm)	Pb	Zn	Cu	Ni	Cr	Rb	Zr	Y	Sr	Ba	Mn
0-10	0	0	2	5	46	7	431	9	55	149	69
10-35	0	0	9	2	41	0	476	7	34	99	65
35-60	0	3	11	8	56	34	422	14	81	236	486

TABLE 4

Per cent elemental analysis of Ca saturated $<2\mu m$ in fraction

Depth (cm)	SiO_2	Al_2O_3	Fe_2O_3	TiO_2	CaO	MgO	Na_2O	K_2O	P_2O_5	BaO	MnO	Loss on ignition
0-10	59.80	19.13	1.84	3.04	0.73	0.22	0.05	1.04	0.13	0.09	0.01	13.70
10-35	72.42	13.00	1.39	5.28	0.34	0.43	0.51	0.80	0.20	0.18	0.02	5.20
35-60	61.92	18.46	4.74	3.42	0.67	0.55	0.25	1.27	0.18	0.07	0.04	8.40

The same can be deduced from CaO values, which indicate the cation exchange capacity (C.E.C.). Calcium ions are added on the clay surfaces by leaching the clay fraction with 4N $CaCl_2$ and subsequently washing with distilled H_2O in order to displace all ions from the exchange surfaces. The Ca ion then reflects the absorption capacity of the clay fraction, i.e., the cation exchange capacity. The measured C.E.C. values are 12.1 meg./100g and 23.9 meg./100g, respectively. X-ray diffractograms of the <2 µm fractions concur with the chemical results (Figs. 7–9). These diffractograms show more

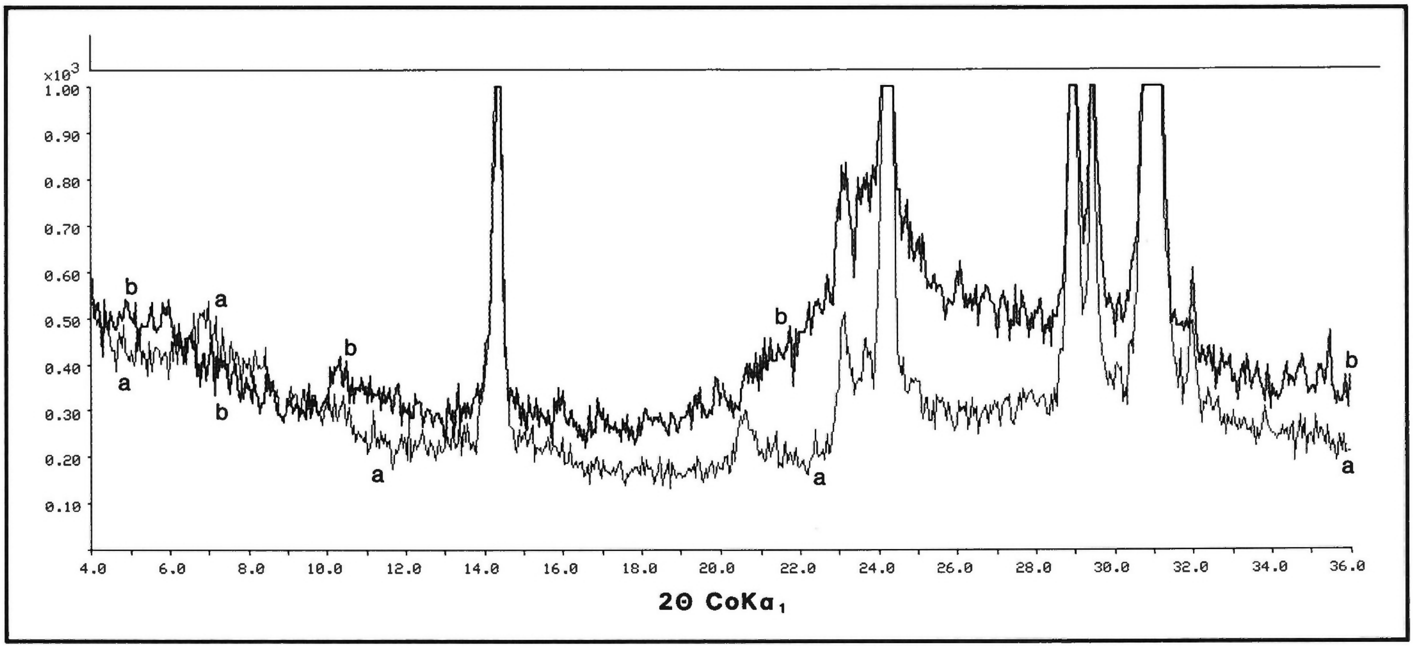

Figure 8. *X-ray diffractogram of the <2 µm fraction from 10-35 cm depth. a, room conditions; b, glycerolation.*

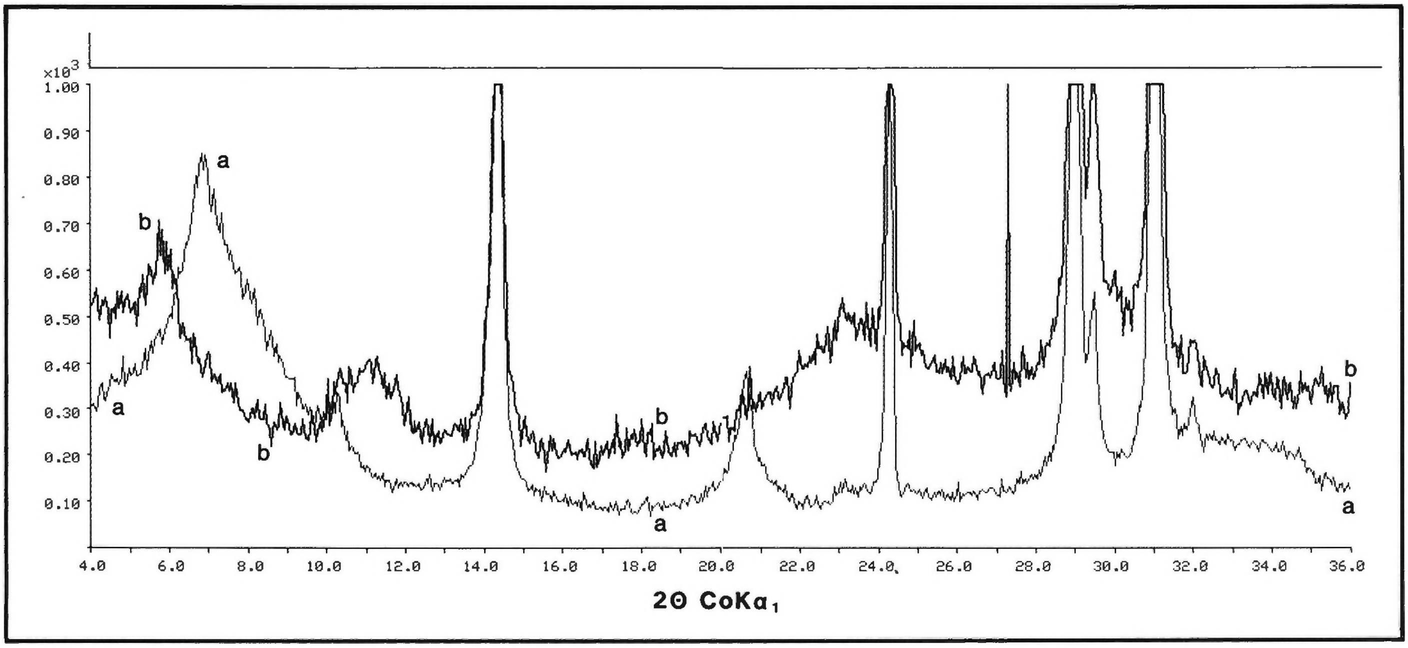

Figure 9. *X-ray diffractogram of the <2 µm fraction from 35-60 cm depth. a, room conditions; b, glycerolation.*

expandable clays and mixed-layer silicates in the upper and lower horizons than in the middle one.

The existence of smectite (d/001 17.9 Å and 17.7 Å) in the upper and lower horizons (Figs. 7,9) and the degradation of smectite (broad hump between 10.1 Å and 17.4 Å with glycerolation; Fig. 8) indicates some weak pedogenetic process, which can be attributed to the acidity of the paleosol. However, the fact that the soil has not been more developed and the horizons are not differentiated adequately under the cover of the conifer forest implies that there was insufficient time to develop this paleosol further. At this stage, the view can be tentatively advanced that this profile is an incipient Gleysol because the horizon below the litter is slightly grey.

Detailed studies on the clay fraction, along with studies of more profiles, will help to clarify whether lack of profile differentiation with these low pH values can be attributed to topography (water table or erosion) or to drastic climatic changes in a very short time.

SUMMARY

Paleosols (Podsols and Gleysols), have been recognized in the Canadian Arctic Tertiary succession (McMillan, 1986; Francis and McMillan, 1987). At the site on Axel Heiberg Island, under the "D" forest litter level, no readily visible soil-forming process has operated. However, the pH is more alkaline lower in the section (Table 2, Fig. 3), indicating the initiation of the soil-forming processes. In addition, the degradation of smectite may be attributed to the initiation of pedogenic processes. The poorly developed profile may be an incipient paleosol.

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THE PALEOLATITUDE OF THE EOCENE FOSSIL FORESTS OF ARCTIC CANADA

E. Irving[1] and P.J. Wynne[1]

Irving, E. and Wynne, P.J., 1991. The paleolatitude of the Eocene fossil forests of Arctic Canada. In Tertiary Fossil Forests of the Geodetic Hills, Axel Heiberg Island, Arctic Archipelago, R.L. Christie and N.J. McMillan (eds.); Geological Survey of Canada, Bulletin 403, p. 209-211.

Abstract

The Eocene "fossil forests" of Axel Heiberg and Ellesmere islands grew between latitudes 74.0°N and 80.2°N, an estimate that includes margins for error.

Résumé

Les "forêts fossiles" d'âge éocène de l'île Axel Heiberg et de l'île d'Ellesmere sont apparues entre 74,0°N et 80,2°N, estimation faite en tenant compte de marges d'erreur.

At present the fossil forests of Axel Heiberg and Ellesmere islands lie between latitudes 77° and 82°N (Christie, 1988). In this study we used paleomagnetic data to determine the latitude at which these forests initially grew, i.e., their latitude in Early to Middle Eocene time, about 45 Ma.

McKenna (1980) considered this question and arrived at an estimate of 78°N. Since his paper was published, paleomagnetic data from Eocene rocks of North America have increased threefold, and although they do not change McKenna's estimate significantly, they allow a more accurate determination to be made.

Relevant North American paleomagnetic data that generously span the age of the fossil forests are given in Table 1. The K/Ar dates and, in one instance, fission-track dates from the rock units studied, lie between 55 and 40 Ma. The seven paleopoles are very well grouped (standard deviation = 4°) and there are no significant differences amongst them. Therefore, it appears that during this time interval the paleopole did not move significantly relative to North America, i.e., latitude did not change.

The paleopole was displaced by 6.6± 1.6° (standard error) from the position of the present geographic pole, in the direc-

TABLE 1

Eocene paleopoles from cratonic North America

Rock unit	Ma	N	Pole (°N, °E)	A_{95}
1. Mistastin Lake impact structure, Labrador	40	10	86,118	3
2. Bearpaw Mountain volcanics, Ma	50-54	30	82,189	6
3. Absaroka volcanics, Wo	48	19	84,177	10
4. Highwood Mountain volcanics, Ma	49-53	30	82,168	7
mean of 1 to 4		4	84,173	5
5. Flores volcanics, B.C.	51±	12	81,188	10
6. Kamloops volcanics, B.C.	49±2	24	81,222	12
7. Kelowna volcanics, B.C.	52±2	28	85,197	10
8. Mean 1 to 7	40-55	7	83,187	3

1. Currie and Larochelle (1969); 2. and 4. Diehl et al. (1983); 3. Shive and Pruss (1977); 5. Irving and Brandon (1990); 6. Symons and Welling (1989); 7. Bardoux and Irving (1989). Ages were derived from K/Ar studies, except 5, which is a U-Pb zircon age. N is the number of sampling sites, A_{95} is the polar error at $P = 0.05$. Data from the Monterey Intrusives of West Virginia, 47 Ma (Lovlie and Opdyke, 1974) are not included because they are based on only 6 sites (see discussion in Diehl et al., 1983). The detailed statistics for entry 8 are 83.4°N, 187.0°E [$K = 397$, A_{63} (standard error) = 1.6°, A_{95} (error $P = 0.05$) = 3.0°].

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[1] Pacific Geoscience Centre, Geological Survey of Canada, P.O. Box 6000, Sidney, British Columbia V8L 4B2

tion of northeastern Siberia. This means that the direction of the north pole, as viewed from the Arctic Islands in the Early and Middle Eocene, was to the west of true north today.

The paleogeographic grid for the Queen Elizabeth Islands (derived from entry 8 of Table 1) is shown in Figure 1, together with fossil forest occurrences. The latitudes of the fossil forest localities range from 75.6° to 78.6°N. Taking errors into account, the localities at the time the forests grew lay between latitudes 74.0° and 80.2°N. This range is about 2° farther south than present latitudes.

Our analysis confirms the estimate given by McKenna, but with additional precision. Evidently the fossil forests grew at very high latitudes, comparable to the latitudes at which they occur today. Undoubtedly, they grew well within the Arctic Circle and were able to withstand extended periods of darkness and sunlight. However, the mean ocean surface temperature at 70°N determined isotopically was 10°C during Eocene time, far higher than today's mean temperature of -10°C (Shackleton and Boersma, 1981; Barron, 1987); the climate was very much milder than at present. This situation, although remarkable, is not unique. For example the abundant *Glossopteris* flora of southeastern Australia flourished within 15° of the south pole in the Permian (Irving, 1964).

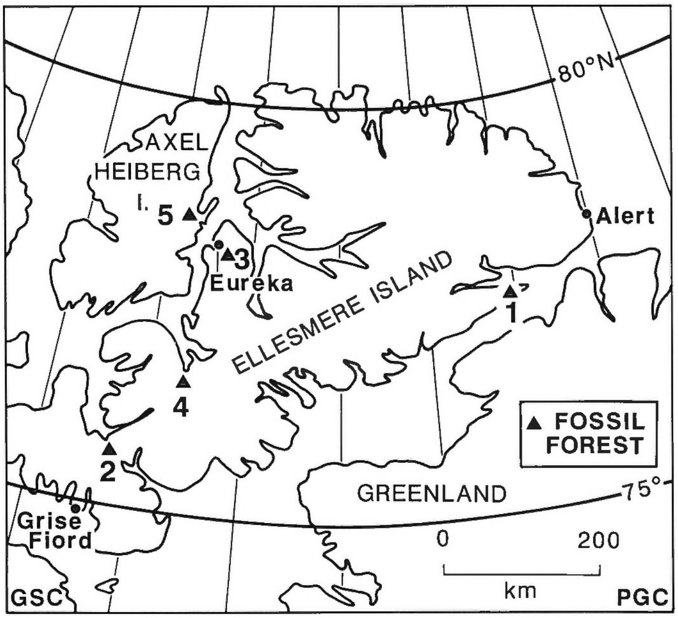

Figure 1. Paleolatitude map of fossil forest localities. The pole of the grid is the mean value, entry 8, from Table 1. The localities, taken from Christie (1988), are: 1. Brainard's "Fossil Forest", 2. Stenkul Fiord, 3. Hot Weather Creek, 4. Strathcona Fiord, 5. Geodetic Hills.

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We would like to acknowledge the helpful review by R.L. Christie.

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THE CONSERVATION OF SPECIMENS FROM THE GEODETIC HILLS FOSSIL FOREST SITE, CANADIAN ARCTIC ARCHIPELAGO

D.W. Grattan[1]

Grattan, D.W, 1991. The conservation of specimens from the Geodetic Hills fossil forest site, Canadian Arctic Archipelago. In Tertiary Fossil Forests of the Geodetic Hills, Axel Heiberg Island, Arctic Archipelago, R.L. Christie and N.J. McMillan (eds.); Geological Survey of Canada, Bulletin 403, p. 213-227.

Abstract

Fossil leaf, cone and wood specimens excavated from the Geodetic Hills fossil forest site on Axel Heiberg Island were found to be very fragile and unstable when brought into the laboratory. Analysis revealed that the wood is preserved in a nonmineralized condition and is very wet to waterlogged. In certain respects, the fossil wood still behaves like recent wood in that it swells or shrinks in damp or dry conditions. It is also compressed. The wood was conserved by a method employing polyethylene glycol 200 and freeze-drying.

Certain cone and leaf specimens are extremely delicate, and experiments demonstrated that the Parylene process could strengthen the specimens, while hardly altering their appearance. It was necessary to pre-treat certain semicoalified samples of leaf mat sections with polyethylene glycol 200 before application of Parylene.

Résumé

On a constaté que les échantillons de feuilles, de cônes et de bois fossiles extraits due site de la forêt fossile des collines Geodetic dans l'île Axel Heiberg étaient très fragiles et instables une fois amenés au laboratoire. L'analyse a révélé que le bois était conservé sous une forme non minéralisée, et qu'il était très humide ou même gorgé d'eau. Â certains égards, le bois fossiles se comporte encore comme un bois récent, du fait qu'il gonfle ou subit un retrait dans des conditions d'humidité ou de sécheresse. Il est aussi comprimé. Le bois a été conservé avec une méthode basée sur l'emploi de polyéthylène glycol 200 et de la lyophilisation.

Certains échantillons de cônes et de feuilles sont extrêmement délicats, et les expériences ont démontré que le procédé basé sur l'emploi du Parylene permettait de consolider les échantillons, tout en modifiant très faiblement leur aspect. Il a été nécessaire de prétraiter certains échantillons semi-houillifiés, constitués de sections de tapis de feuilles, avec due polyéthylène glycol 200 avant l'application de Parylene.

INTRODUCTION

The specimens collected from the Geodetic Hills fossil forest site on Axel Heiberg Island, N.W.T. are, unlike most fossils, very unstable. This paper describes conservation treatments that enable such specimens to be exhibited and curated with less risk of damage.

The problem was brought to our attention in late 1986, when a large fossil tree stump excavated early in the summer at the fossil forest arrived at the National Museum of Natural Sciences in Ottawa. The stump is shown in Figure 1, after treatment. It was apparent that the stump was very unstable; even though wrapped in plastic film and damp newspaper, it had begun to crack and delaminate. The museum requested the assistance of the Canadian Conservation Institute in the hope of saving the collected stump. Examination of the wood showed that it was much wetter than had been initially supposed, and, although the fossil material is about 40 million years old, enough cellulose remains to cause significant shrinking upon loss of moisture.

The fossil forest site also yielded an unparalleled number of unique and beautifully intact leaf and cone fossils. Many proved to be very fragile, and were very difficult to conserve using conventional methods. This hindered the assembly of collections and made museum display almost impossible. Furthermore, the extreme delicacy of some specimens limited the scope of laboratory investigation by paleobotanists. In 1986, there were no methods available to preserve the specimens.

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[1] Canadian Conservation Institute, Department of Communications, 1030 Innes Road, Ottawa, Ontario K1A 0C8

The goal of conservation is to preserve specimens with minimal intervention so that their appearance and nature are altered as little as possible. To carry out a conservation treatment, it is therefore necessary to understand the composition and the nature of the material to be treated, with special regard to deterioration. It is usual to carry out sufficient analysis to obtain an understanding of the behaviour of the materials and an indication of the responses to treatment. For the wood, these investigations have characterized the major alterations that took place after the death of the tree and have addressed the problem of shrinkage after excavation. For the leaf and cone fossils, the main problem is their extreme fragility.

FOSSIL WOOD

The wood remains of the fossil forest occur in a range of sizes and conditions. Essentially all parts of the original trees have survived burial and the wood may be found as stumps or as pieces of branch or trunk wood that vary from a few centimetres to over a metre in length. A notable feature of the fossil wood is the very clean breaks, resembling brash breaks characteristic of rotten wood. The wood is dark brown to almost black and has a resinous odour. Amber or resin is noticeable as small golden brown granules, present in scattered pockets.

The wood from the fossil forest is some of the oldest known that remains in a nonmineralized condition. Analysis is therefore intrinsically interesting and may help to reveal the reasons for the remarkable preservation; however, properties considered relevant to conservation are of particular interest. Factors to be examined include: those related to dimensional stability, such as the moisture content and the shrinkage behaviour; and those relevant to conservation treatment, such as relative density and the content of cellulose, lignin and ash.

Wood from two stumps was studied. The first, NMNS S1 (recovered by J.F. Basinger, University of Saskatchewan, in 1986), is a complete stump measuring about one metre in height and about the same in width at the buttress level. After excavation and some drying, it separated into three large sections, which were labelled A, B and C.

Obst et al. (*this volume*) and Young (*this volume*) report NMNS S1 A as a species of conifer, but definitely not *Metasequoia*. Young (*this volume*), however, has identified HR4 as probably *Metasequoia*. Section NMNS S1 A was used in our experimental work. A small outer portion (about 20 cm x 10 cm x 1 cm), which had separated from the trunk, was used for the analytical investigation.

The second sample, labelled HR4, was from a stump found on the east side of fossil forest hill. Its heartwood was absent; all that remained was a ring-shaped fragment that presumably delineated the former outer surface (see Fig. 2). This trunk had been excavated in sections and the locations of each recorded so that the trunk could be reassembled. The stump was divided into a number of approximately equally sized

Figure 1. *Part of a large fossil tree stump after treatment with polyethylene glycol (PEG) 200.*

samples suitable for carrying out experimental trials. Sections around the "ring" were cut at intervals of 10 cm. Length was typically 15 cm, and thickness was either approximately 2 cm or 8 cm.

Moisture content

Percentage moisture content is calculated as 100 times the weight of water in the wood divided by the oven dry (i.e., completely dry) weight of the wood. Oven drying of samples from the stump at 110°C gave initial moisture values of approximately 24 per cent *after* some drying had occurred subsequent to its excavation. Field measurements were carried out with a Delmhorst resistance moisture meter on a partially buried trunk at a site on the "East Hill" designated "Carya Fruit Level A". (This meter has two kinds of probes. A two-electrode, insulated probe allows the measurement of moisture at depths of up to 3 cm below the surface of the wood, whereas a four-electrode probe measures surface moisture content only.)

The trunk measured 1.8 m in length, 13 cm in thickness and 20 cm in width. About 40 cm of trunk was exposed, and the rest remained buried. At the deepest point, the trunk was 33 cm below the ground surface. The moisture readings are shown in Table 2.

It is evident from this investigation that buried wood gives moisture readings in excess of 30 per cent. Exposed wood has a much lower moisture content, and there is an abrupt transition at ground level to the higher readings.

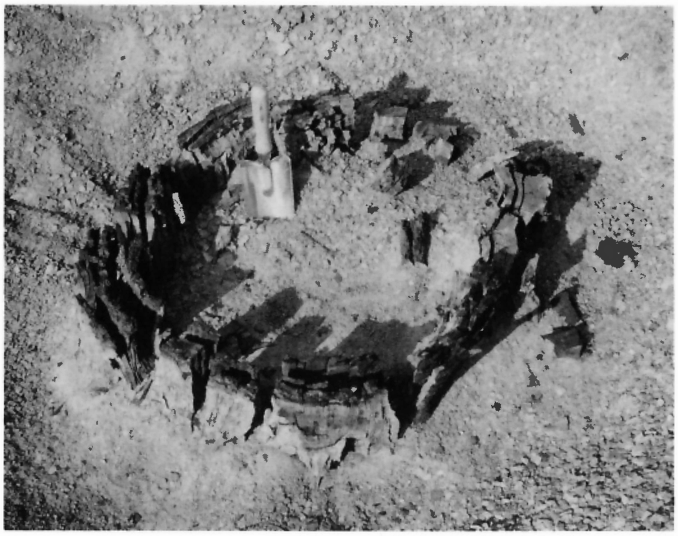

Figure 2. *A typical example of stump with heartwood absent. All that remains is a ring-shaped fragment that presumably delineated the outer surface.*

Shrinkage

Cubes from NMNS S1 A (about 2 cm on each side) were dried completely. They were then weighed, measured, and subsequently re-humidified at 100 per cent rh (relative humidity) until they achieved the fibre saturation point[2]. Moisture absorption was monitored by following weight change. It was assumed that once constant weight was attained the samples would be at equilibrium. This took approximately 200 hours. They were then weighed and measured. Samples were placed in desiccators at 0 per cent rh to achieve complete dryness, after which measurements were repeated. Samples were then re-humidified to constant weight, and all measurements were repeated again.

The experiment showed that loss and regain of moisture, shrinkage, and swelling were reversible processes and that in this respect the fossil wood behaves like new wood. Overall volumetric shrinkage was 11.0 per cent. Table 1 compares these data with figures for new wood.

The experiment also showed that the fibre saturation point for the fossil wood is 18.9 per cent.

TABLE 1

Percent shrinkage and swelling of fossil wood

	Average of three specimens	Average of swelling and shrinkage	New[a] wood *(Sequoia sempervirens)*
Longitudinal			
swelling	3.81%	3.86%	(~0.5%)
shrinkage	3.90%		
Radial			
swelling	4.48%	4.14%	2.6%
shrinkage	3.79%		
Tangential			
swelling	4.59%	4.21%	4.4%
shrinkage	3.82%		
Volumetric			
swelling	11.12%	11.04%	6.8%
shrinkage	10.96%		

[a] From Summitt and Sliker, 1980.

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[2] At fibre saturation, all of the cell wall in normal wood is fully saturated with water, but the cell cavities are empty. As a result, the wood occupies its maximum swollen volume. This occurs when wood is exposed to 100 per cent relative humidity. For normal wood, the moisture content is usually about 30 per cent at this point.

Chemical composition

Ash

The ash content, which is an indication of the mineral component of the wood, was measured for four samples of NMNS S1 A. The procedure involves heating at 600°C following ASTM[3] D1102-56.

The results were: 1.08 ± 0.04 per cent per unit weight, or 0.78 ± 0.01 per cent per unit volume. The ash was white and had a fibrous texture.

Extractives

Hot-water solubility was carried out according to the procedure given by ASTM D110-56. This analysis is normally done to estimate tannins, gums, sugars, proteins, inorganic soluble salts, and starch. It was anticipated that the fossil wood would have little soluble material. 2.16 per cent of the sample was extracted with hot water.

Lignin

The lignin content was measured on four samples of NMNS S1 A by the procedure given by ASTM D1106-56. The average lignin component was 71 ± 10 per cent of the nonextracted oven-dry wood.

Cellulose

The analysis for holo-cellulose[4] was performed on only one (1.6 g) sample. The procedure followed ASTM D1104-56, which isolates pure holo-cellulose. Holo-cellulose was measured at 28 per cent for one sample of NMNS S1 A. The product was perfectly white and gave an infra-red spectrum that closely resembles that of modern cellulose (Whatman filter paper A; see Fig. 3).

An attempt to go one step further and isolate alpha-cellulose with the procedure described in ASTM 1103-60 failed to yield any product. It is assumed that, because of the degraded state (as described below), the test cannot distinguish degraded alpha-cellulose from hemicellulose.

Attempts were made to measure the degree of polymerization (D.P.) of cellulose by measuring the viscosity of a solution in cadoxen. (This standard procedure is described in Browning, 1967). D.P. was evaluated at 145.

Collaborative work on other aspects of the wood has been carried out at the Canadian Conservation Institute (see Young, *this volume*). Chemical analysis of some of the samples has also been conducted at the U.S. Forest Products Laboratory (see Obst et al., *this volume*).

Relative density

Relative density is defined as the weight of wood substance in the oven-dry condition divided by the volume of the sample in the fully water-saturated or wet condition (that is, the volume occupied by the specimen at the fibre saturation point). Relative density is expressed as kg/m^3.

For NMNS S1 A, the average relative density obtained for six samples was 719 ± 35 kg/m^3. For HR4, relative density, as determined from three samples of approximately 150 g each, was 870 ± 10 kg/m^3.

TABLE 2

Field moisture readings[1]

Distance above or below ground	Distance along trunk (cm)	Meter reading: Four-electrode probe	Meter reading: Two electrode probe
+9	0	8.5	—
+5	15	8.0	—
0	30	37.0	—
-9	60	34.0	—
-19	90	35.0	—
-26	120	40.0	58.0
-33	150	40.0	—

[1] expressed as percentage of oven-dry weight of wood.

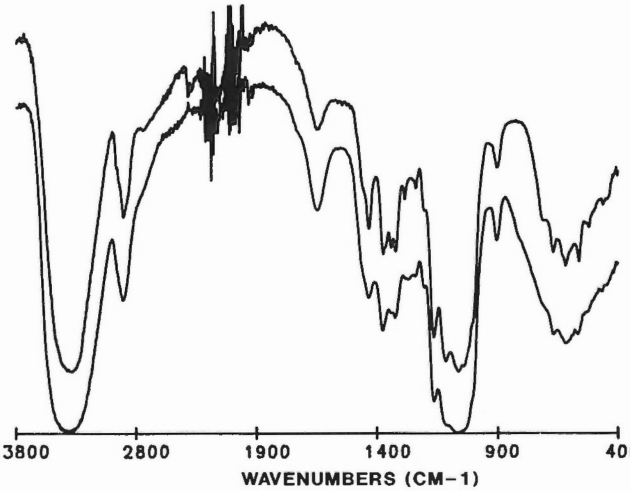

Figure 3. *Infra-red spectra of recent and ancient cellulose. The upper curve is the spectrum of alpha-cellulose (Whatman Grade A filter paper), and the lower is that of holo-cellulose extracted from the fossil wood.*

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[3] The American Society for Testing and Materials (ASTM) sets American National Standards.

[4] Holo-cellulose is a term used to describe total cellulose, i.e., hemicelluloses and alpha-cellulose.

Discussion

Moisture content

In laboratory tests, moisture meter readings of the fossil wood were within two or three per cent of the laboratory calculated values. The readings are thus a rough guide to actual values. The moisture content of buried wood is well above the fibre saturation point (i.e., maximum swollen state). Allowing for the fact that the fossil wood is severely crushed, the wood evidently approaches a waterlogged condition. The wood when buried therefore remains intact because it remains swollen; exposed wood, however, fragments and delaminates as it shrinks. The wood, on excavation, is well above equilibrium moisture content and is thus in an unstable state.

Shrinkage

The shrinkage of the fossil wood is quite different from that of normal wood, as shown in Table 1. It is interesting that the fossil wood has lost the anisotropic behaviour of fresh wood, which shrinks much more across, rather than along, the grain. In normal wood, shrinkage arises principally from the loss of water from the secondary cell wall, which is composed mainly of cellulose. This isotropic shrinkage may well be a consequence of the shortness of the cellulose polymer chains and the consequential loss of intermediate fibril structure. What is important for conservation is that the shrinkage is quite high, and much higher than for normal wood, along the grain. Measures need to be taken to eliminate or reduce such shrinkage.

Chemical composition

Chemical analysis of the wood (summarized in Table 3) clearly shows that the bulk of the wood substance is lignin. This is to be expected because lignin is the most stable component of wood. Obst et al. (*this volume*) examine the degree of methylation and show that the lignin is in a relatively undeteriorated state.

The relative proportions of lignin and cellulose in the ancient wood are very different from those of new wood (see Table 3). The proportion of cellulose in the fossil wood is much reduced. In this respect, the fossil wood is comparable to old (1000- to 3000-year-old) waterlogged archaeological wood previously examined by Grattan and Mathias (1986), a typical analysis of which is also included in Table 3. This similarity suggests a similar history: immersion in anaerobic bog water for a very long time. Because of this similarity we were able to use a conservation procedure that has been successful for archaeological wood.

Old waterlogged wood usually has a rather high ash content: 2 per cent or more, and frequently as high as 8 to 10 per cent. The ash content of new wood is often found to be about 0.2 to 0.5 per cent. The ash content of 1.1 per cent for the fossil wood from Axel Heiberg Island is below normal for waterlogged wood and above normal for new wood. This clearly shows that no significant mineral replacement has occurred. The absence of significant ash also suggests that the fossil wood should be easy to impregnate.

The D.P. of the cellulose at 145 is much less than that of cellulose from present day trees, which varies between 5000 and 10 000 (Panshin and de Zeeuw, 1970). It is thus not surprising that an attempt to separate the alpha-cellulose component failed to yield any product.

The shortness of the cellulose chains means that the tensile strength of the wood is severely reduced. Stresses set up by shrinkage cause the wood to break readily, and this accounts for the fragility of the wood when exposed to the elements. It also emphasizes the need to eliminate shrinkage to prevent disintegration of wood in storage or display.

Relative density

Wood from NMNS S1 A has a much higher relative density than would be expected for deteriorated wood. (For example, uncrushed deteriorated *Sequoia*, with a chemical analysis matching that of the fossil wood, is estimated to have a relative density of between 100 and 200 kg/m^3.) The reason for this discrepancy is clearly that the fossil wood has suffered severe crushing. Microscopy of HR4 (see Young, *this volume*) shows that crushing is greater in the horizontal plane of the wood than in the vertical. Young (*this volume*) also notes that the early wood was much more severely crushed than the late wood and that crushing had taken the form of a lateral shear. The ray cells passing through the late wood were at an angle of 43 degrees to those in the early wood.

The degree of crushing was also estimated from the weight of residual lignin per unit volume (510 kg/m^3) calculated from the chemical analysis. Assuming that residual lignin represents half of the original (by analogy with more recent water-

TABLE 3

Chemical composition of the fossil wood (NMNS S1 A)

Chemical component	Fossil wood Amount present (%)	Fossil wood Cumulative total (%)	New wood *Picea* sp.	Archaeological wood (%)
Lignin	71.0	71.0	26.3	60.7
Holo-cellulose	28.1	99.1	67.9	19.4
Ash	1.1	100.2	0.2	3.7
Hot water extract	2.2	02.4	3.7	4.1

logged wood), the original lignin density would be 2 x 510 = 1020 kg/m^3. Since lignin would probably have represented about 30 per cent of the original dry wood substance, if no loss of material had taken place, the relative density of the wood in the crushed state would be 100/30 x 1020 = 3400 kg/m^3. If the original relative density of the fresh wood is estimated to have been approximately 400 kg/m^3, the wood must have been crushed by a factor of 3400/400 = 8.5.

Another method of estimating the amount of crushing is to examine branches and trunks, most of which are elliptical in cross-section. Presumably, the longer axis of the ellipse represents the original width and the minor axis indicates the results of crushing by downward pressure. If it is assumed that lateral spreading is negligible and that a cross-section of the original wood was roughly circular, the ratio of the major and minor axes should reveal the amount of crushing. A typical horizontal timber (Carya fruit level B) was 17 cm wide and 1.5 to 2.0 cm thick. The ratios of major axes to minor (and hence estimated crushing) range from 8.5:1 to 11.0:1. By contrast, the timber used in the moisture measurement experiment described above was 20 cm wide and 13 cm thick, which gives a crushing ratio of only 1.5:1. The degree of crushing obviously varies from place to place within the fossil forest.

The second implication of the high relative density and the crushing is that the pore space (or cell cavity volume) is less than in normal wood. This will result in reduced accessibility to the interior of the wood and reduced space available for accommodating a conserving impregnant, compared to uncrushed wood. Pore space is estimated as follows:

If we assume, from Barbour (1984), that cell wall substance has a density of about 1500 kg/m^3, then cell wall substance in the fossil wood (relative density 719 kg/m^3 for NMNS S1 A) occupies 719/1500 x 100 = 47.9 per cent of the volume of the wood. The remaining volume (52 per cent) is pore space, and this space would contain 520 kg of water. For NMNS S1 A, maximum moisture content when the pore space is full of water (i.e., fully waterlogged) is about 520/719 x 100 = 72 per cent.

With normal wood, the air space is much greater. *Sequoia*, for instance, with a relative density of 400 kg/m^3, has 73 per cent pore space, which means that the moisture content when fully waterlogged is 183 per cent[5].

In field measurements (see Table 2), the moisture meter indicated levels of between 37 and 58 per cent, i.e., well above the fibre saturation point. Moisture meters are notoriously inaccurate at these high levels. It therefore seems likely that the fossil wood in its present buried condition approaches a fully waterlogged state with a moisture content of 72 per cent.

Stabilization experiments

Slow drying

Stump NMNS S1 A had dried significantly while in transit from the field, and the drying process was allowed to continue under more controlled conditions. The rate of moisture loss was controlled by wrapping a polyethylene sheet around the sample. Moisture content readings were taken regularly with permanently installed electrodes (Grattan, 1989). Moisture was recorded in five locations and observed over 311 days. In about 15 days, 50 per cent of the moisture had been lost, whereas after 100 days, 75 per cent had gone. About 300 days were required to reach equilibrium. Final moisture content was 7.1 ± 0.5 per cent. Slow drying is a technique that can, in some cases (where there is little chance of collapse), be used to stabilize waterlogged wood (Grattan and Clarke, 1987). Unfortunately, despite the slow, careful drying, shrinkage was quite substantial in the fossil forest wood and much cracking and warping resulted. It was clear that more experimental work was necessary, and it was decided to treat the wood with polyethylene glycol (PEG) and to freeze-dry it, as is done with archaeological wood.

The PEG freeze-drying treatment

The similarities between the fossil wood and archaeological, waterlogged wood invite attempts to control the shrinkage and cracking using techniques that are most effective for such wood (Grattan, 1988). Freeze-drying after impregnation with PEG prevents shrinkage, can strengthen wood, and is effective on dense and impermeable wood. It does not cause drastic changes in appearance of the surface, unlike older or more conventional procedures. Amorphous regions of the cell wall are penetrated with PEG, which behaves chemically similar to water. The PEG is absorbed predominantly by cellulose. It was believed that the fossil wood would act in much the same way because it contained recognizable cellulose. Because the wood changed in volume as moisture was absorbed or desorbed, it seemed reasonable to suppose that PEG impregnation would help to eliminate shrinkage.

An experiment was carried out to determine the effects on the fossil wood. Low molecular weight PEGs were selected because cell wall penetration of impregnants is slow, even in new wood, and diffusion rates through bulk wood were also likely to be very slow because the fossil woods are compressed. In the normal procedure, wood is soaked in an aqueous PEG solution of gradually increasing strength until the

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[5] Moisture content for wood is calculated as noted earlier under that heading. The weight of water in wood can exceed that of the wood; the moisture content of waterlogged wood consequently is greater than one hundred per cent.

water within the wood has been completely exchanged with the solution. The wood is then frozen at -20°C or below and subjected to vacuum sublimation.

A calculation for the treatment of waterlogged archaeological wood was used to determine the most suitable solution strength (Grattan, 1988). The correct impregnating solution contains enough PEG to fill the second order space (the microcapillary network within the cell wall) at the completion of freeze-drying. Test samples were placed in various solutions to explore the optimum conditions. Thick and thinner sections were tested in order to determine approximate diffusion rates. Two grades of PEG, 200 and 400, as well as di-ethylene glycol were investigated to see if diffusion was significantly faster with an impregnant of lower molecular weight.

Experimental method

Wood was taken from sample HR4. Samples were weighed and immersed in tap water. Because no micro-organism formation was noted, it was not necessary to add biocides.

The outlines of the samples were traced on waterproof paper to record approximate initial dimensions. More accurate measurements of dimensional change were provided for by the installation of three small brass screws. These allowed for measurement along the grain (longitudinal) and across the grain (tangential). It was not possible to install screws so that radial changes could be measured.

Five wood samples were allowed to dry without treatment for three months at 50 per cent rh and room temperature until they achieved constant weight. The shrinkage values for these control samples were used to estimate the effectiveness of the PEG methods in controlling shrinkage (see Table 4). The average value was used in the calculation of anti-shrink efficiency (A.S.E.)[6].

Three treatment agents were tested: PEG 200, 400 and di-ethylene glycol. Impregnant concentrations (i.e., final level achieved) of 10, 20 and 29 per cent were used in order to investigate the relationship of glycol concentration to the shrinkage.

Progress of impregnation was monitored by periodic weighing of the samples fully immersed in the treatment solution. This was accomplished by suspending samples by fine wires below a balance. Typical increases in weight are shown in Figure 4. Diffusion of di-ethylene glycol was slightly more rapid than that of PEG 200, which was much more rapid than PEG 400. For small samples of wood, impregnation typically reached 95 per cent of the finally attained level within approximately 20 days for di-ethylene glycol, 30 days for PEG 400 and 45 days for PEG 400. Impregnation was continued until weight change had almost ceased.

Results

The effectiveness of treatment was monitored by measurement of A.S.E. A plot of A.S.E. versus final concentration of impregnant is shown in Figure 5. The point at 100 per cent A.S.E. gives the theoretical PEG concentration necessary to fill the residual cell wall completely with PEG. This estimate is based on a calculation of residual cellulose and has been described elsewhere (Grattan, 1988). Results for the small samples fall close to the straight line drawn between the point at 100 per cent and the origin. For the smaller samples, the best A.S.E. was obtained with PEGs 200 and 400. Di-ethylene glycol was consistently less effective by 5 to 15 per cent. Values for the larger samples, denoted by L on the graph, with one exception, fall well below the line, suggesting that impregnation was not complete.

The appearance of the treated wood was only slightly altered: surfaces were not much darker than untreated ones. The wood was still very fragile and had a very thin, delicate surface layer, which tended to flake off.

Recommendations

PEG 200 at a concentration of 39 volume per cent is recommended as the most suitable impregnant for HR4, which has a density of 870 kg/m^3. The most suitable concentration for other specimens varies with wood density. Large specimens require impregnation periods of up to two or more years. (The longest impregnation period used in the experiments reported here was 362 days. This was clearly insufficient for the larger samples.) Impregnation may be carried out

TABLE 4

Shrinkage values for control samples (%)

	Longitudinal	Tangential
HR4/2/F	8.94	5.84
HR4/2/G	6.75	5.17
HR4/1/B2	7.1	5.32
HR4/3/C	7.51	6.51
HR4/3/D	7.4	7.0
Average	7.54	5.97

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[6] Anti-shrink efficiency is one hundred times the percentage of shrinkage after treatment divided by the percentage of shrinkage after uncontrolled air drying. An anti-shrink efficiency of 100 per cent means that all shrinkage has been prevented, whereas 0 per cent means that the shrinkage is the same as if uncontrolled air drying has taken place.

in one stage because the danger of cellular collapse is slight for this wood. Freeze-drying should be carried out by precooling the wood to -40°C for at least 24 hours. (Longer periods may be necessary for larger samples.) The wood should then be kept in the drying chamber at -20°C until all moisture is removed.

After the PEG freeze-drying treatment, further consolidation may be necessary because of the surface fragility noted above. This could be achieved in a number of ways, depending on specimen size. For small items, the Parylene[7] method, described below for cones, gives excellent results, but the method is not feasible for whole stumps because of the limited size of the coating apparatus. For large pieces, the more conventional methods using acrylic resins, such as Rohm and Haas Acryloid B72, or polyvinyl butyral resins, such as Monsanto Butvar B98, could be used. However, these consolidants are best avoided because they cause drastic changes in appearance.

Treatment of stump NMNS S1 A

With the investigation of PEG complete, attention was turned to stump NMNS S1 A. Before impregnation could proceed, it was necessary to return the material to a wet condition. This was effected in a plastic enclosure in which the

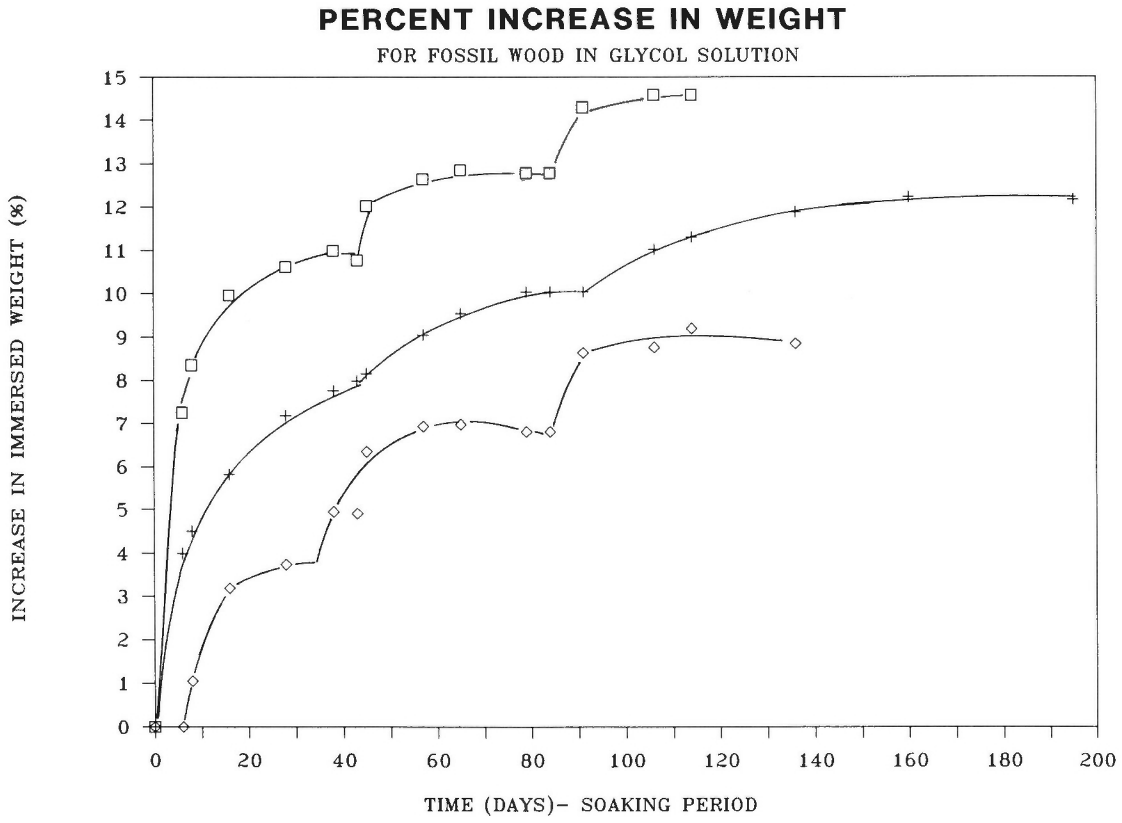

Figure 4. *Percentage change in sample weight, measured with samples fully immersed in PEG impregnation solutions. Increases in solution concentration cause rapid weight change as new solution is absorbed. Each sample was treated first in 10 per cent w/v glycol, then in 20 per cent w/v and finally in 29 per cent w/v. Each curve thus shows three distinct phases. Each phase has an initial rapid weight increase, followed by a period in which weight gain gradually decreases and then stops. When weight gain ceased, the sample was placed in a solution of the next concentration level or was removed for freeze-drying, if it had achieved final concentration. Illustrated are curves for small samples in which rapid diffusion took place. □ designates HR4 (2B), which was treated in PEG 200. + designates HR4 (2A), which was treated in PEG 400. ◇ designates HR4 (2D), which was treated in DEG.*

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[7] A trade name of the Union Carbide Corporation.

relative humidity was gradually increased to 100 per cent. As earlier, electrodes were used to monitor moisture content. After 160 days the stump reached equilibrium and was then suspended in a tank of water so it could be weighed while immersed. After the weight had stabilized (after 30 days), the water was exchanged for a 30 volume per cent solution of PEG 200. Weight readings were taken at weekly intervals. After nine months of increase, weight had stabilized and the stump was prepared for freeze-drying. This was carried out under vacuum, and the stump is now dry and stable. It is somewhat cracked, but wax resin mixtures and silicone rubber have been used to immobilize the cracks.

CONES AND SEEDS

Sample description

Cones of species *Metasequoia*, *Taxodium*, *Glyptostrobus*, *Pinus*, *Picea* and seeds of *Carya* were considered together as a single conservation problem.

Some cones are flattened and are structurally strong (although they have brittle appendages), whereas others are less compressed and are generally delicate. Figure 6 shows a comparison between a recent (1987) cone of *Metasequoia* and a fossil cone of the same species from the fossil forest. Figure 7 shows one of the more fragile cones. Although most cones have, like the wood, little or no mineral content, some are extensively mineralized. As is the wood, the cones are found in a wet condition, and as they dry the less compressed ones tend to open. In doing this they mimic the behaviour of recent cones, which open as they ripen. As the cones open (especially the *Picea*), they become extremely fragile. In this condition, they are very difficult to handle: the gentlest touch with a camel hair brush can easily dislodge or fracture the scales.

Although no analytical studies were carried out, the cones appear to be mainly composed of ligneous material. Little deterioration of the ligneous fraction is evident in the fossil wood and, by analogy, the cones are probably in a similar condition.

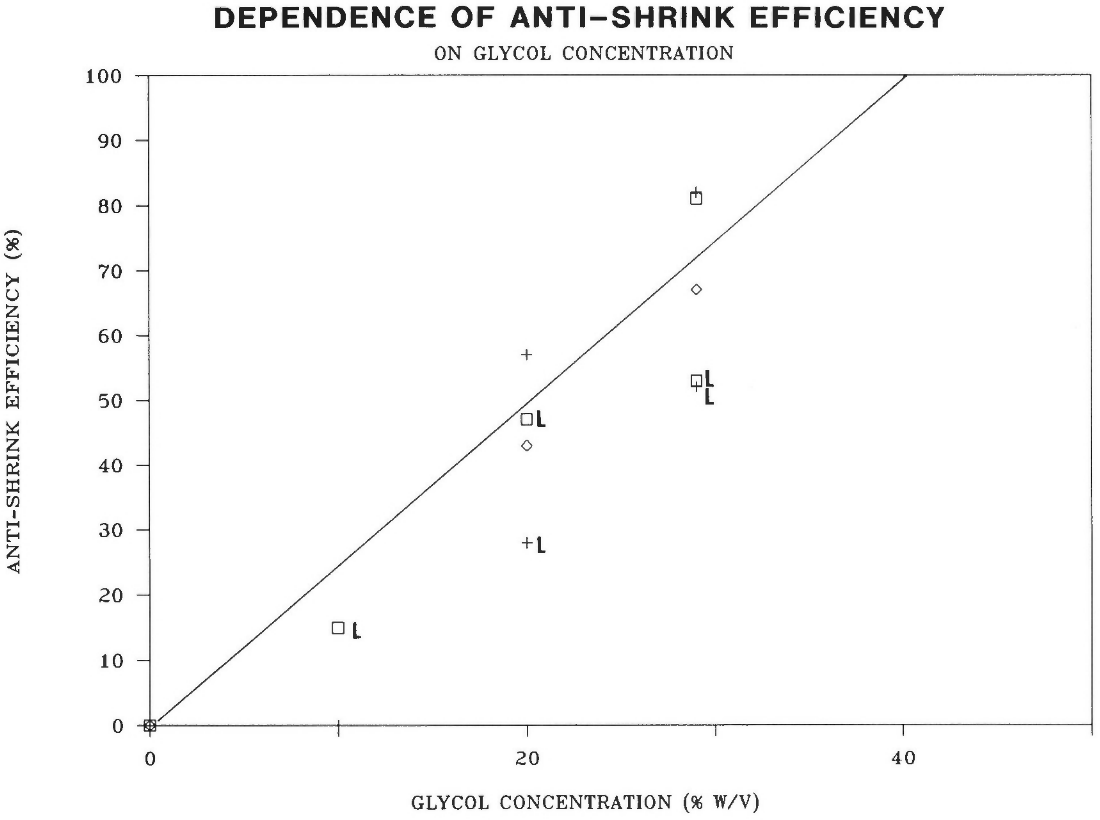

Figure 5. *Anti-shrink efficiency (A.S.E.) versus the concentration of PEG solution used for impregnation. The A.S.E. values are the average of the cross-grain (tangential) and longitudinal measurements. Large samples are denoted by L. Point labelling: PEG 400 treatments are shown by* □, *PEG 200 by* + *and DEG by* ◇. *The line is given by theoretical considerations. The value at 100 per cent is the calculated solution strength necessary to prevent all shrinkage (i.e., A.S.E. = 100 per cent).*

The fragility of the cones rules out normal conservation procedures. Most are too delicate to survive application of consolidant by brushes, and some cones would not survive the surface tension forces were they to be immersed in a solution of a resin. Even if brushing on, or immersion in, resin were a possibility, the application of resin to their porous surfaces would cause a drastic darkening.

A method of strengthening in which consolidant is introduced in the gaseous phase seemed most promising. However, very few methods have been developed that are capable of doing this, and of these, only the Parylene method has been developed for conservation purposes.

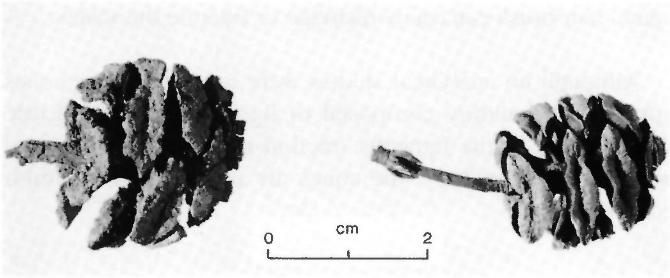

Figure 6. *A comparison between a recent (1987) cone of* ***Metasequoia*** *(right) and a fossil cone of the same species from the fossil forest (left).*

Figure 7. *An example of a particularly fragile cone after drying in the laboratory.*

Parylene treatment

Poly(*p*-xylylene), or Parylene, was discovered in the 1940s. Although its unique properties have been known for many years, it was not practical for museum application until Nova Tran Inc., a Union Carbide subsidiary, developed suitable equipment to carry out the process in a controlled fashion. The process, which takes place under vacuum, is one in which the active monomer of Parylene is deposited on all exposed surfaces, whether internal or external. The active monomer then polymerizes in situ to encapsulate the specimen in a perfectly conforming coating of a linear aromatic polymer.

The method begins with the sublimation of a cyclic dimer of *p*-xylylene, which is then passed through a pyrolysis tube at 670 ± 20°C. At the high temperature, the dimer splits to create two active monomer units. These are allowed to diffuse through the pyrolysis tube into the sample chamber, which is at room temperature. Upon collision with a surface, the active monomer units condense, then polymerize with other condensing monomer units to form a coherent polymer film that conforms perfectly to the surface of the substrate (Fig. 8). The monomer molecules, being small, are able to penetrate porous substrates so that more than the outer surface of a specimen is coated. The thickness of the Parylene coating can be controlled to within a micron or two, and film thickness, depending on the requirements, commonly ranges from 1 to 20 µm.

Two versions of Parylene are commonly used: "N" is a pure hydrocarbon, whereas "C" has one chloride on each aromatic ring. Parylene "N" is used in applications where better diffusion is required. The Parylene method has one limiting feature as a conservation technique: the finality of the process. Once applied, Parylene cannot be safely removed.

The impracticality of removal of the conserving material poses certain ethical problems. One of the principles of conservation is that treatments should be reversible, i.e., it should be possible to return the object to its pre-treatment condition. With Parylene this is not possible, so its use must be carefully considered in terms of 1) the lack of alternatives, 2) the rate of deterioration of the specimen in the absence of treatment, 3) the stability of Parylene and the permanence of the treated object, and 4) the requirements for use of the specimen.

Experimental method

Initial experiments were carried out with a small group of cones. The cones selected included four very deteriorated *Picea*, four rather flattened and less delicate *Metasequoia*, and four *Picea* in an intermediate state of preservation. A critical part of the process is the initial evacuation of the sample chamber: for successful operation, the pressure must be reduced to below 20 millitorr[8]. Outgassing, i.e., loss of moisture, interferes with Parylene deposition and must be complete before coating can commence. The cones were watched closely during the evacuation procedure. There was some opening of scales, but not as much as in the initial drying. During the deposition of Parylene, no change in appearance of the cones was observed. On removal of the cones, it was apparent that the 12.8 µm deposit of Parylene C had considerably strengthened them.

Examination under the stereo microscope at x100 magnification showed little evidence of Parylene. This "invisibility" arises from the unique conforming nature of the coating. In thin section, however, the Parylene is clearly visible as a uniform, birefringent outer layer.

A second experiment was devoted to the problem of cones that were too fragile to remove from the surrounding deposit. A large *Picea* cone that was too delicate to lift from its surroundings was placed, with its matrix, in a small tray inside the coating chamber. As a first step, a thin (<1 µm) layer of Parylene N was applied. When this had diffused into the cone and its matrix, the cone was removed from the surroundings, cleaned and given a 7.8 µm coating of Parylene C for final consolidation. The thin layer of "N" did not prevent removal of particulate matter that had adhered to the cone due to the Parylene coating. The main problem in the procedure is the scattering of very light particles of the organic matrix inside the chamber.

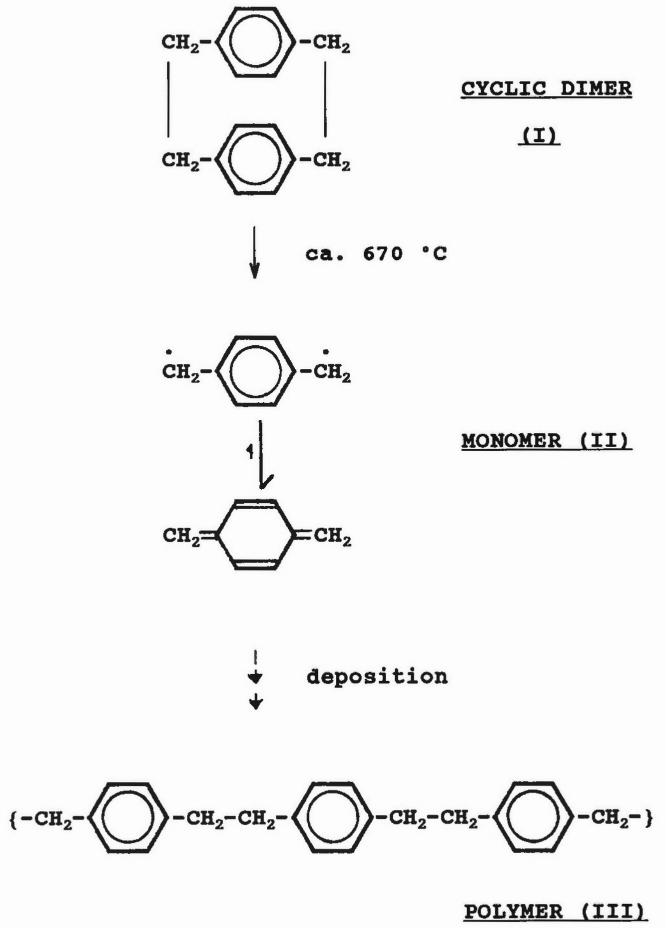

Figure 8. *The formation of Parylene.*

Most cones were successfully consolidated by application of from 10 to 12 µm of Parylene C; however, it was sometimes found that the attachment of scales was poor. This was particularly noticeable in a large *Picea* cone.

In a third experiment, initial coating with Parylene N was increased to 3.6 µm and final coating with Parylene C to 13.7 µm, to give a total thickness of 17.3 µm. Attachment of scales was much better at this thickness. In later work with the cones, the initial coating of Parylene N was found to be unnecessary.

A group of cones treated with Parylene is shown in Figure 9.

Following a suggestion by B.A. LePage (pers. comm.), as a precursor to Parylene treatment, some of the stronger but flattened and mineralized cones were soaked in various concentrations of hydrofluoric acid. This was followed by neutralization with sodium carbonate solution and washing in distilled water. The acid removed siliceous and other mineral accretions. This process was particularly useful for partially mineralized cones in which the morphology was obscured. It was found that treatment with 12 per cent solution for six hours provided the most aggressive conditions that could be tolerated without damage to the material, especially where cones were more heavily mineralized. After drying, the scales on flattened cones tended to open, giving a better impression of the original cone. The acid treatment, however, darkened the cones considerably. The acid-treated cones were then successfully consolidated with 18 to 21 µm coatings of Parylene C.

Treatment of the museum collection

The Parylene method was used to treat a cone collection assembled by Dr. L. Hills for the National Museum of Natural Sciences. All the cones were catalogued according to species, integrity, and condition. Specimens with good display potential were treated with 20 µm of Parylene C to give them maximum strength for endurance. The remainder of the collection, primarily intended for study, was divided into two groups, with representatives of all cone types in each. One group,

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[8] The Nova Tran Inc. machine is calibrated in these units of pressure. One torr = 1 mm of mercury column.

about two thirds of the material, was treated with 10 µm of Parylene C, while the second group was left untreated. The 10 µm coating was thought to be the minimum that would allow some handling, while permitting microscopic examination. The untreated collection was retained in case long-term stability problems with the Parylene should appear, or in case it is discovered that certain studies are impeded by the layer of Parylene.

Discussion

Initial results of Parylene treatment are considered to be entirely satisfactory. Consolidation was effective at a Parylene thickness of 12.8 µm or greater. For cones with large scales, it was found useful to treat first with Parylene N to diffuse consolidant into the body of the cone as far as possible, and then to add Parylene C to give a combined thickness of about 20 µm. Such thick coatings are essential for cones in which the scales have small attachment points. Because of the conforming nature of the Parylene layer, the coating is equally thick regardless of the cross-sectional thickness of the fossil substrate.

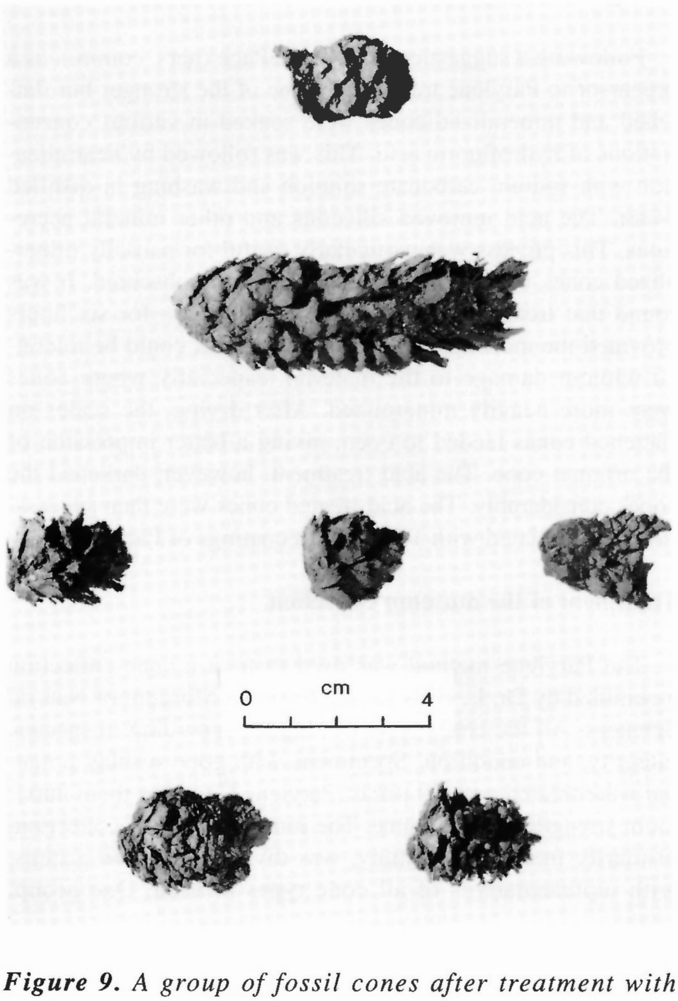

***Figure 9.** A group of fossil cones after treatment with Parylene.*

The Parylene treatment of the cones can be justified because of the lack of effective alternatives and the acute need to strengthen disintegrating material. Because it cannot be known what method of study or analysis may be required in the future, it seemed wise to leave a representative selection of cones uncoated, even though this may result in their (eventual) physical disintegration. As for Parylene itself, studies have shown that it is very stable, although it probably is affected by UV light (Grattan, 1990).

Changes in the appearance of the cones have been impossible to quantify by instruments that measure colour and gloss because of the fragility (before coating) and the irregularity of the surfaces. It is therefore relevant and interesting to examine some of the data acquired with other materials that have been coated with Parylene. Colour measurement of leaf mat sections (described in the next section), before and after Parylene coating, shows that colour is little altered by the process. It is reasonable to suppose that this is also true for the cones.

FOSSIL LEAF MATS

Sample description

"Leaf mat" is the term used to describe the fossil remains of the forest floor. Leaf mats are mostly composed of compressed leaves, but they also include wood fragments, seeds, cones, and other plant parts. The major component is leaf litter. The dominant species is *Metasequoia*, but many others have been found including *Larix*, *Picea*, *Pinus*, *Glyptostrobus*, *Taxodium* and *Platanus*. No chemical or physical analyses of the leaf mats were done. However, mats appear to vary considerably in condition from highly compressed, semicoalified deposits in which individual leaves are not easily distinguished, to noncompressed, noncoalified layers with each leaf and cone separate and distinct. As with the cones, the leaf mats become excessively fragile on drying. In the field, they have but an ephemeral existence: they simply blow away as dust as they are exposed and dry out.

Four leaf mat samples, representing different states, were studied. Each required a different conservation procedure. The same basic procedures used for the cones were used here.

Experimental method

Experiment 1

Sample 11 (28 cm x 18 cm x 3 cm) from the *Glyptostrobus* Horizon, on the East Hill Site, Geodetic Hills (collected by L. Hills in 1988) was treated. This specimen was described by Hills as a "good example of weathered *Metasequoia* leaves for museum display". Although noncoalified and composed of easily separable *Metasequoia* leaves, it was quite well consolidated. It was treated with 8.7 µm of Parylene N to obtain some penetration. No dimensional change was observed and the Parylene treatment gave a completely satisfactory consolidation. Colour measurements were made at seven points on the sample, both before and after treatment. A Minolta Chroma Meter[9] was used following the CIE (Commission International de l'Éclairage) standard under C illumination conditions.

The values given in Table 5 are the average of the seven measurements. They indicate a slightly less reflective (i.e., darker) surface, with a very slight trend toward blue, after treatment. The measurements confirm visual assessment that almost no colour change had occurred.

Experiment 2

A second, more fragile, leaf mat sample, number 12 (20 cm x 13 cm x 4 cm), was also treated. This was less compressed and contained much more distinct and separate leaves, including those of *Metasequoia* and *Larix*. Because it was also rather thick, we hoped to split it horizontally to produce more surfaces for display. It was coated with 4 µm of Parylene N; this light coating allowed the mat sample to be handled and separated, although it remained very fragile. The thickness of the sample increased from 4 to 6 cm on evacuation of the vacuum chamber.

The sample was separated using spatulas and by sliding in sheets of paper. The upper layer received a second coating of 14 µm of Parylene C, which consolidated the material satisfactorily and still allowed the removal of individual leaves. Again, appearance was little altered. Figure 10 shows the condition after treatment.

The lower layer of the sample was coated with 30 µm of Parylene C. Consolidation was excellent, and, even with such a thick coating, appearance was little altered. With this thickness, however, the sample had become in effect a solid block, and separation of individual leaves was impossible.

TABLE 5

C.I.E. Colour co-ordinates

	X	x	y
Sample 11			
Before	11.58	0.3365	0.3402
After	7.37	0.3301	0.3321
Sample 12			
Before	9.90	0.3321	0.3347
After	5.61	0.3330	0.3347

Figure 10. *A sample of leaf mat after treatment with Parylene.*

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[9] A trade name for this instrument.

Colour measurements were made at six points on the upper layer of the sample, both before and after treatment. A Minolta Chroma Meter was used following the CIE standard with the C standard illuminant (see Table 5).

As in the previous example, the decrease in reflectance (X) indicates darker shades. The colour change, almost negligible, is nearly imperceptible to the eye.

Experiment 3

A third experiment was carried out with a 25 cm x 15 cm x 4 cm piece cut from sample 4, a rather coalified specimen. Unfortunately, woody parts were substantially distorted and the sample tended to fragment on vacuum drying within the Parylene coating chamber.

Experiment 4

The fragmentation and warping of the coalified leaf mat, sample 4, described in Experiment 3 was a very unsatisfactory result. A second piece of sample 4 was, therefore, treated in the same manner as was the fossil wood, as described earlier. A 37 cm x 10 cm x 21 cm piece was placed in a 30 per cent w/v solution of PEG 200 for 3.5 months. It was then chilled to -40°C and freeze-dried at -20°C. No warping took place, and it was then coated with 26 µm of Parylene N. The Parylene consolidation was completely satisfactory, and no sign of delamination or separation of the Parylene film was detected. It did, however, take a very long time (about 100 hours) to coat.

Experiment 5

A sample of leaf mat (sample 3) of a condition intermediate between coalified and noncoalified, which had been kept wet since excavation, was treated by Parylene alone. As in Experiment 3, the results were very poor and a second experimental treatment was tried with a 28 cm x 25 cm x 9 cm portion of the sample. Unlike sample 4, sample 3 was too fragile to withstand impregnation in PEG solution without support; it would have disintegrated. The section was therefore tightly sewn into a lightweight synthetic mesh bag, the mesh allowing the PEG solution to penetrate. Treatment followed that described in Experiment 4, except that the Parylene was applied in two steps: 4 µm of Parylene N followed by 30 µm of Parylene C. Use of Parylene C, which is deposited much faster than Parylene N, greatly reduced the time required to coat the section. (The Parylene C application took only four hours.) Sufficient Parylene N had been added to penetrate and to give some internal strengthening. Again, the treatment was successful, and there was no obvious incompatibility between PEG and Parylene.

Discussion

Parylene works well in the consolidation of fossil leaf mat. Coatings can be applied thinly enough so that leaves may be strengthened but are still separable one from another. Thicker coatings can cement a block completely, but with much alteration in appearance. The variation between leaf mats is such that general rules for treatment are not useful; the conditions of treatment must be appropriate to the state of each sample.

CONCLUSIONS

Conservation procedures for wood, leaf, and cone specimens from the fossil forest at the Geodetic Hills site were developed. The experiments with Parylene are very encouraging: the problem of how to consolidate fragile material without sacrificing appearance appears to have been solved. Use of PEG prior to Parylene deposition is of benefit to samples susceptible to shrinkage and disintegration. No incompatibility problems were observed.

ACKNOWLEDGMENTS

I thank Dr. S. Cumbaa, Mr. R. Day, and The National Museum of Natural Sciences for assistance and support in the field; Dr. T. Austin and Mr. B. Humphrey and the Union Carbide Corporation for practical assistance, advice, and the loan of a Parylene coater; Ms. S. Drouin for carrying out chemical separations on the fossil wood; Mr. C. Gruchy for support and advice; Dr. L. Hills for practical help in the field and for selecting fossil samples suitable for conservation; and Mr. C. McCawley for his unstinting support of laboratory and fieldwork. The Polar Continental Shelf Project, Department of Energy, Mines and Resources, supported the fieldwork.

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ERRATA

p. 148: The first line after the heading "Petrology" should read
"The coals (<20% ..... "

p. 149: In Table 1, the Ro values should read as follows:
Sample No. 10-15-85 —— 0.38; Sample No. 2-20-85 —— 0.37;
Sample No. 3-20-85 —— 0.45

p. 150: In the caption for Figure 2, "b, cutinite" should read "b, sporinite", and "c, sporinite" should read, "c, cutinite"

p. 156: Figure 1 should read "Massive resinite (R) showing evidence of degradation."