A PALYNOLOGICAL INVESTIGATION OF THE LATE QUATERNARY VEGETATIONAL HISTORY OF THE BAIE VERTE PENINSULA, NORTHCENTRAL NEWFOUNDLAND

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BY

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ABSTRACT

A Palynological Investigation of the Late Quaternary Vegetational History of the Baie Verte Peninsula, Northcentral Newfoundland

Key Words: *Pollen analysis, vegetation reconstruction, postglacial, climatic change, deglaciation, Newfoundland, species migration.*

Palynological studies of peat and lacustrine sediments in Atlantic Canada have provided outlines of the late- and postglacial development of the vegetation. Furthermore, these vegetational changes have permitted researchers to interpret late-glacial and Holocene climatic changes. On the Island of Newfoundland, however, the relative paucity of palynological work has hampered the development of a regional synthesis of vegetational and climatic history. This thesis examines the sequence of vegetational changes in northcentral Newfoundland in order to add to the knowledge of late Quaternary palaeoenvironments of this province.

Duplicate cores from two lakes on the Baie Verte Peninsula were obtained for pollen analysis, radiocarbon dating, and loss-on-ignition analysis. The results indicate that the Peninsula was ice-covered during the Late Wisconsin and that the ice limit probably extended at least to the northern terminus. The northern highlands were deglaciated by 11,800 BP and dissipation of the ice progressed by downwasting and ice-recession toward the interior of the peninsula.

Pollen percentage, concentration and influx diagrams are presented and a tentative regional pollen zonation for the northern Baie Verte Peninsula is proposed as follows: (I) Gramineae-herb zone, before 11,800 BP; (IIa) *Betula*-Cyperaceae subzone, 11,800 to 10,000 BP; (IIb) Shrubs-*Picea* subzone,

10,000 to 8500 BP; (IIIa) *Alnus-Abies* subzone, 8500 to 6700 BP; (IIIb) *Betula-Picea-Alnus* subzone, 6700 to 3200 BP; (IV) *Picea-Betula-Alnus* zone, 3200 BP to present.

Pioneering communities were replaced by a dwarf-shrub tundra after 11,800 BP. A prolonged shrub-tundra phase was probably a result of residual ice in lowland areas which acted as a physical barrier to tree migration and which may have affected the local climate. Arboreal immigration commenced after 9500 BP with *Picea*. The closing of the forest canopy and the development of a white birch-black spruce forest occurred after 6700 BP. From that time to about 3200 BP the climate was warmer and drier than at present. A subsequent cooling trend is inferred primarily from a resurgence of black spruce at the expense of white birch. A significant decline in total pollen abundance after 2000 BP is attributed to a decrease in the density of the regional vegetation and decreased sediment focussing.

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CHAPTER 1 INTRODUCTION

Pleistocene glaciations resulted in the displacement of plant species and the eradication of former vegetation patterns in eastern North America. Following retreat of the Late Wisconsin ice and the emergence of isostatically depressed areas, recolonization and subsequent changes in the development of vegetation led to the contemporary location of vegetation zones and present ranges of plant taxa.

During the past two decades, an increasing number of studies have been undertaken concerning the process of revegetation in eastern Canada. Questions pertaining to the extent of Late Wisconsin ice, the pattern and chronology of glacial recession, locations of plant refugia, the rate and direction of plant migration as well as the composition of early forests, the role of different plant species in postglacial communities and interpretations of the palaeoclimate have been tackled. Gradually, the sequence of vegetational and climatic changes is being revealed.

For the mainland portion of eastern Canada, the emphasis in palaeoecological studies is shifting. Now that the Holocene pollen sequence for the region has, to a large extent, been established, current research is concerned increasingly with multiple site analysis, palaeoclimate modelling and plant population dynamics. For example, the compilation of pollen data from multiple sites has enabled the preparation of isochrone maps depicting the spatial and temporal movement of ecotones and specific taxa (e.g. Bernabo & Webb 1977; Webb et al. 1983). Similar data are being used to produce isotherm maps for specific time periods during the Holocene (e.g. Bartlein & Webb 1985). In addition, interpretations of fine resolution pollen records have begun to provide detailed information on plant population and community dynamics, particularly the role of fire in forest ecosystems, and on sedimentological processes (e.g. Green 1981,1982,1983).

In contrast, palynological studies in Newfoundland are still in the stage of reconnaissance insofar as large gaps in the palaeoenvironmental history of the province need to be filled. Although studies under way in the northcentral and central parts of the island (Macpherson in prep.) are rapidly correcting this situation, many basic questions remain.

With this regional imbalance in mind, the present project was designed to elicit information on the Late Quaternary environments of an area of Newfoundland. It was decided that the pollen analysis of two lake sites from the Baie Verte Peninsula could provide essential data on palaeoenvironments since the glacial history of this area remains speculative and its vegetational and climatic history has not previously been studied.

As a background to this study, a review of the known and speculative ice limits and ice-free areas in Newfoundland is given in Chapter 2. This chapter also discusses plant refugia hypotheses since an implicit aim of this study was to determine whether the Baie Verte Peninsula could have been a refugium for plants. In addition, the late- and postglacial sequence of vegetation for the Atlantic region and Newfoundland is discussed as well as the inferred climatic history of Newfoundland for the same time period.

The field work, laboratory procedures and pollen analytical methods employed in the project are discussed in Chapter 4. This precedes a discussion of the physical environment of the Baie Verte Peninsula (Chapter 5) and the characteristics of each of the two sites investigated (Chapter 6). From the pollen data obtained from the sites, a series of pollen zones were identified and are described in Chapter 7. Sediment description and results of radiocarbon dates and loss-on-ignition analysis are also presented in this chapter.

This information provides the basis for Chapter 8: the interpretation of the pollen and stratigraphic data. Taking each site in turn, the chapter discusses deglaciation and provides a reconstruction of the vegetation and landscape using the approaches outlined in Chapter 4. In addition, the migrational routes of certain taxa are considered in Chapter 8. The final section of this chapter attempts to intepret the climatic history from the inferred vegetation sequences.

CHAPTER 2 BACKGROUND AND LITERATURE REVIEW

2.1 Introduction

In order to describe the late- and postglacial vegetational development of a region, it is essential to establish what is known of Late Pleistocene environments. This is the aim of this chapter. The Late Wisconsin ice extent in the Atlantic region is outlined with emphasis placed on known and speculative ice margins and ice-free areas in Newfoundland. A brief assessment of plant refugial hypotheses is followed by a discussion of deglaciation and sea-level changes in the Atlantic region. Revegetation of the eastern Canadian seaboard is the focus of the fifth section followed by a review of palynological research in Newfoundland including both palaeovegetation and palaeoclimatic reconstructions.

2.2 Late Wisconsin Ice Extent

[i] The Atlantic Region

The extent of the last or Late Wisconsin glaciation in the Atlantic region has been the subject of considerable debate over the past century. The large body of literature amassed over this period has been reviewed by several workers. Grant (1977b) and Ives (1978) discussed the stages of Quaternary investigation in the Atlantic region, and Tucker (1976), Brookes (1982) and Rogerson (1982, 1983) have reviewed interpretations of the glacial history of Newfoundland.

From 1860 to 1940 the prevailing view among researchers (e.g. Daly 1902; Coleman 1920, 1926; Fernald 1925) was of limited ice along the eastern seaboard and the persistence of ice-free areas within the ice limits of the last glaciation. The work of Flint (1940) heralded a major shift in viewpoint and it became the accepted position that all high coastal summits were ice covered (Ives 1978).

Flint's (1940) work formed the basis of the 'maximum' model of Late Wisconsin ice cover. In the 'maximum' reconstruction, Late Wisconsin Laurentide ice extended to Long Island, New York. The ice overran the Laurentian Channel and the Bay of Fundy, coalescing with local ice caps over Nova Scotia, Prince Edward Island and the island of Newfoundland, and extended southeast of Nova Scotia to a moraine system on the Scotian shelf (Mayewski et al. 1981).

Since about 1970 the earlier view has regained approval based mainly on the work of Grant (1969a,b, 1972) in Newfoundland. His research on moraine complexes and ice-flow indicators pointed to a radiating island-based ice cap on Newfoundland (Mayewski et al. 1981) and the inundation of only the northern tip of the Northern Peninsula with Labradorean ice.

These studies, together with more recent research (e.g. Prest et al. 1976; Brookes 1977a; Grant 1977a,b; Tucker & McCann 1980), have formed the basis of the 'minimum' model of Late Wisconsin ice cover. In this reconstruction, Laurentide ice did not extend beyond the North Shore of the Gulf of the St. Lawrence and restricted ice caps covered the island of Newfoundland, New Brunswick and Nova Scotia while the Laurentian Channel, Bay of Fundy, the Magdalen Islands, Anticosti Island, the eastern portion of Prince Edward Island and certain coastal areas of Newfoundland and Nova Scotia remained ice-free (Mayewski et al. 1981). A recent study by Quinlan and Beaumont (1982) shows that the two extreme models of Late Wisconsin ice cover in the Atlantic region produce hypothetical postglacial sea levels that bracket actual observations of sea levels. They therefore conclude that Late Wisconsin ice distribution must have been intermediate between the two proposed reconstructions.

Figure 2.1 shows the inferred limits of Late Wisconsin ice in the Atlantic region as proposed by Grant (1977b). Grant's suggestion of a restricted terrestrial ice cover is mainly derived from research on weathering zones in mountainous terrain (Brookes 1977a; Grant 1977a) in Newfoundland (see below). Grant's restricted limit of Late Wisconsin ice in the Laurentian Channel is controversial. He bases this reconstruction on a lack of evidence of Late Wisconsin ice on the Magdalen Islands or western Cape Breton Island. A recent study on Anticosti Island (Painchaud et al. 1984), however, indicates that surface till was deposited during the Late Wisconsin readvance of the Laurentide ice sheet. As the evidence points to the ice limit as being maintained near the south coast of Anticosti Island, the Esquiman Channel may also have been inundated.

[ii] Newfoundland

The ensuing discussion outlines the known and speculative limits of Late Wisconsin ice in Newfoundland. Two reconstructions have been advanced for the Northern Peninsula. One implies that Labradorean ice impinged upon the lower parts of the extremity of the peninsula (Grant 1977b), whereas the other implies a complete separation of the ice sheet over Labrador from the ice cover of Newfoundland (Rogerson 1982; Prest 1984). The latter reconstruction is based on the proposition that ice-free conditions existed in southeast Labrador, east of the Paradise Moraine (Fulton & Hodgson 1979).

Both models, however, imply that the main source of ice in the Northern Peninsula was from the Long Range Mountains. This is supported by ice flow patterns in the northern uplands (Waitt 1981) and the pattern of moraines in lowland areas (Grant 1969a,b). Similarly, to the south, piedmont glaciers issued

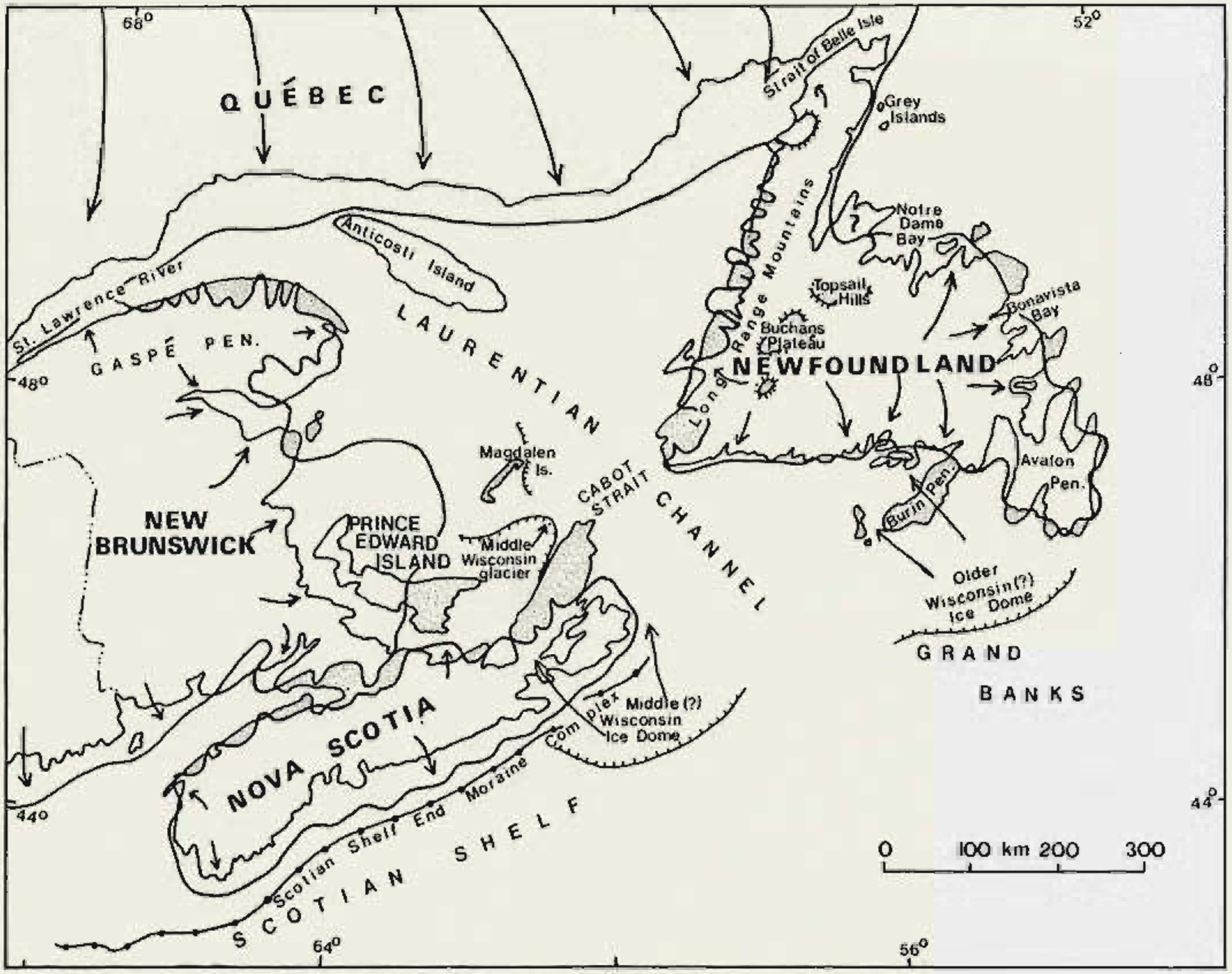


Figure 2.1: Speculative limits of Late Wisconsin ice, nunataks (stippled), patterns of ice flow (arrows), and limits of earlier ice masses (pecked lines).
(redrawn from Grant 1977b).

from troughs in the Long Range Mountains (Grant 1969b) and Anguille Mountains (Brookes 1975), depositing moraines along the coast.

In these mountain ranges, Grant (1977a) and Brookes (1977a) have identified three distinct altitudinal weathering zones, that is, "discrete areas exhibiting different degrees of weathering" (Grant 1977b, p.255). The youngest weathering zone, the upper limit of which is 600m, is characterized by a freshly striated surface mantled with erratics. Above this zone, extensive nunataks are believed to have existed (Grant 1977a). Grant (1977a) proposed that summits in the area above 750m have been ice-free since before the last interglacial based on the relative depth of fluvial dissection of the surface and the mantle of felsenmeer. However, other workers (e.g. Mayewski <u>et al.</u> 1981) have challenged the hypothesis that felsenmeer is indicative of ice-free conditions. They argue that if the base of the ice sheet were frozen to the substrate, previously weathered debris would be afforded protection and left intact after deglaciation thereby wrongly implying non-glacial conditions.

Although Grant's (1977b) reconstruction of the ice maximum shows the limit arching across Notre Dame Bay leaving terminal parts of the Baie Verte and Bonavista peninsulas ice-free, no firm evidence has been found to determine the seaward extent of ice in northcentral Newfoundland. Earlier workers in the area (Jenness 1960; Lundqvist 1965) argued that an inner-outer drift zone boundary, separating ground moraine on the coastal side from younger eskers, kames and ground moraine on the inner side, resulted from either a still-stand during the Late Wisconsin or a final major advance. Tucker (1973, 1974) took issue with the latter interpretation. His work on the Halls Bay Deltas indicated that a calving front rapidly deglaciated the bay until the ice became land fast in valleys to the south. More recently, marine cores retrieved about 50km offshore, in Notre Dame Channel, reveal a continuous sequence of postglacial sediments overlying unstratified glacial till (Mudie & Guilbault 1982). A stratified till overlying the unstratified till has a radiocarbon age of ca. 18,000 BP and Mudie and Guilbault postulate that the basal unsorted till marks the presence of grounded ice during the maximum extent of the Late Wisconsin glaciation.

On the south coast of Newfoundland, the Late Wisconsin ice margin is generally envisaged as extending to just beyond the present coastline. A study of the Burin Peninsula (Tucker & McCann 1980) describes the limit of the main Newfoundland ice cap as a well-defined feature lying across the Gisborne Basin, slightly north of the limit proposed by Grant (1977b), and points to the existence of small separate ice caps along the central rib of the upper peninsula. More recently, Leckie and McCann (1982) provided evidence pertaining to the pattern of ice in the Hermitage Bay area. They claim the northern part of the area was glaciated by the main Newfoundland ice cap and the southern part by a small, upland ice cap, broken by nunataks.

The sequence of glacial events in central Newfoundland is unclear due to the lack of dated material. According to Vandeveer and Sparkes (1982), glacial flow indicators suggest that the area was glaciated during the Late Wisconsin. Earlier papers by Grant (1977a,b) agree in general with this reconstruction, but suggest that the Topsail Hills, Buchans Plateau and Annieopsquotch Mountains may have lain above the ice cover.

The Avalon Peninsula had its own ice cap that discharged into Placentia

Bay and Trinity Bay where it merged with ice from central Newfoundland (Henderson 1972). Both Rogerson and Tucker (1972) and Mellars (1981) have attempted to delimit the seaward extent of the Avalon ice cap. Tracts of felsenmeer on the Bay de Verde (Carbonear) Peninsula (Henderson 1972) imply ice-free conditions in that area, and Grant (1977b) proposed that the extremities of other peninsulas of the Avalon may also have lain beyond the Late Wisconsin ice limit. However, Mellars' attempt to determine the extent of glacier ice on the St. John's Peninsula proved unsuccessful. The earlier workers, concentrating on moraines and raised marine landforms in bays and on islands, proposed that Late Wisconsin ice extended beyond the eastern coast. How far the ice overran the continental shelf remains uncertain. However, research into the provenance and transport of marine sediments in the Atlantic Region (Piper & Slatt 1977) indicates that terrestrial sediments were deposited on the inner continental shelf east of Newfoundland. This, at least, suggests that the Grand Banks lay beyond the ice maximum.

During the last glaciation, the island of Newfoundland supported multiple ice caps. Grant (1974) identified the sites of several local remnant ice caps. Rogerson (1981) argued that these sites may not have supported stagnating ice but, rather, supported active dispersal centres during the Late Wisconsin maximum. According to Rogerson (1982), these dispersal centres consist of an ice divide along the northern peninsula; a radial outflow centre north of Red Indian Lake, a second west of Meelpaeg Lake, and a third in east-central Newfoundland; and a complex of centres over the Avalon Peninsula comprising a central ice cap and ice divides along the Bay de Verde and St. John's Peninsulas. Grant (1974) supported the idea that Late Wisconsin ice caps were also centred on the Bonavista, Baie Verte and Burin Peninsulas, but Rogerson (1982) claims that the two latter were the sites of pre-Late Wisconsin dispersal centres.

2.3 Deglaciation and Sea-Level Changes

[i] Atlantic region

Radiocarbon dates from terrestrial deposits and marine shells have provided a basis for the reconstruction of the overall pattern of ice retreat in the Atlantic region. In general, deglaciation occurred before 13,000 BP in the south along the Bay of Fundy coast and as late as 10,500 BP in southeastern Labrador, although Newfoundland does not fit neatly into this sequence (see below). A conflict in the chronology for the North Shore of the Gulf of St. Lawrence is apparent. Dubois and Dionne (1982) claim that the coastline was ice-free by ca. 10,500 BP. Other evidence, however, indicates that part of the coast, near Sept-Iles, was deglaciated ca. 1000 years later (Lowden et al 1971). This date is in closer agreement with the timing of deglaciation of northern Anticosti Island ca. 9500 BP (Painchaud et al 1984).

[ii] Newfoundland

In Newfoundland, the pattern of Late Wisconsin deglaciation is difficult to discern because of the presence of several ice caps rather than a single ice sheet (Rogerson 1982) and the paucity of reliable dates. In coastal areas, a portion of the evidence is derived from dated marine material. Macpherson (1982a) argues, however, that if indeed several coastal areas lay beyond the ice margin, as has been postulated (e.g. Rogerson 1981), then marine shell dates may not directly reflect the timing of deglaciation. Carefully selected terrestrial sites may, on the other hand, provide dates more indicative of deglaciation (see Figure 2.2).

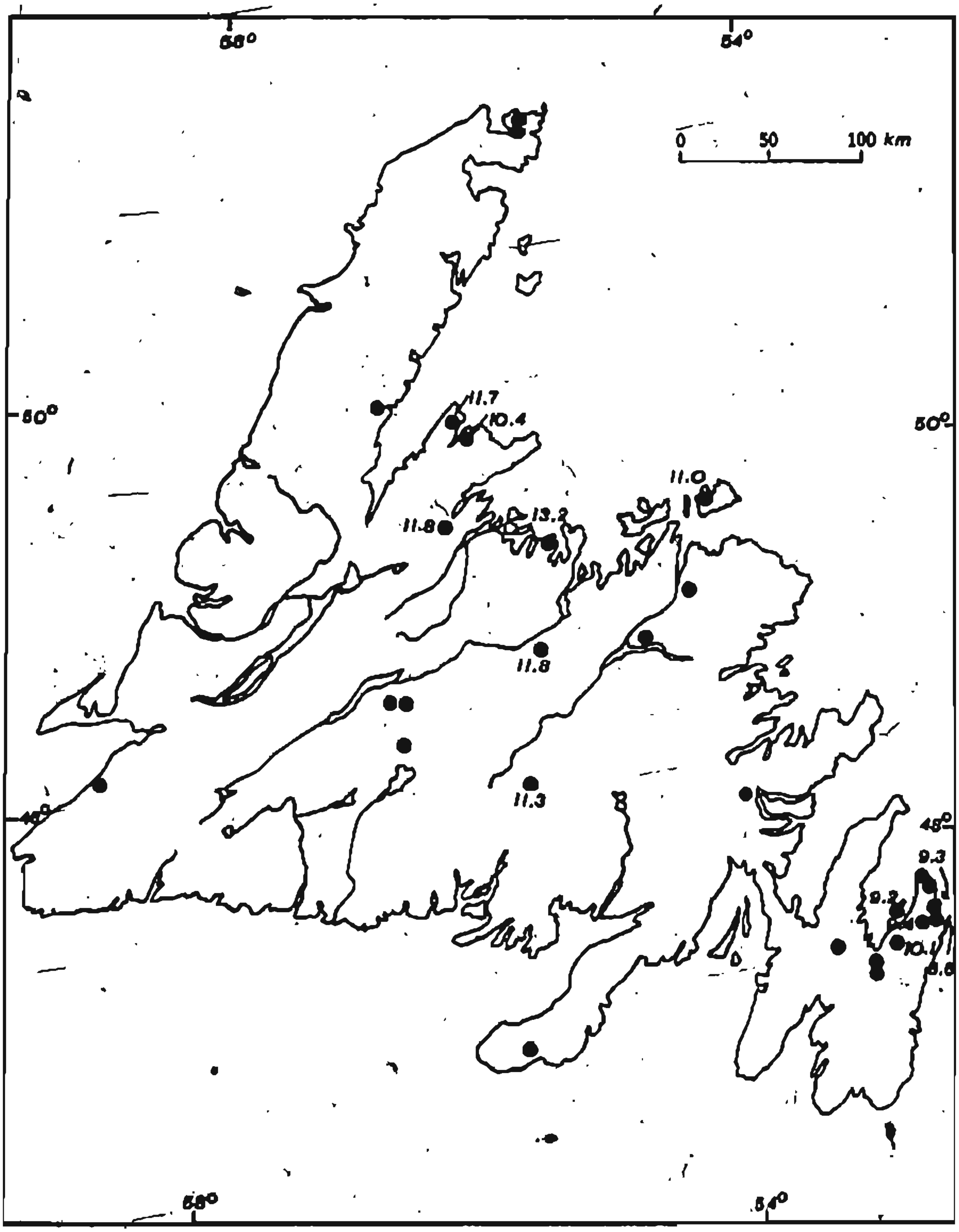


Figure 2.2: Location of pollen study sites including those with dated basal material (in x10³ Yrs BP) indicative of deglaciation.

(Sources given in text)

A discussion of ice retreat in Newfoundland follows. On the southwest coast, near Cabot Strait, dated marine shells overlying a till provide a minimum age of 13,800 BP for the deglaciation and submergence of the area (Brookes 1975). On the west coast, in St. Georges and Port au Port Bays (locations shown in Figure 2.3), deglaciation occurred slightly later: marine shell dates from deltas place the event between 13,200 and 13,600 BP (Brookes 1969). In St. Georges Bay, a glacial readvance occurred ca. 12,600 BP, producing the Robinson Head moraine (Brookes 1977b). Farther north, dated shells from glaciomarine rythmites place deglaciation of the Bay of Islands at ca. 12,600 BP (Brookes 1977b). Near the tip of the Northern Peninsula, at Ten Mile Lake, a major ice advance occurred ca. 10,900 BP (Grant 1969a) although the moraine could be interpreted as the Late Wisconsin maximum (Rogerson 1982, 1983).

Recently obtained dates from basal lake sediments in northcentral Newfoundland have provided indispensable information for reconstructing the pattern of deglaciation in that area. A basal date of 11,800 BP from a site in the southern part of the Baie Verte Peninsula (Macpherson in prep.) accords with the date of deglaciation at the head of Halls Bay, where marine shell dates indicate an ice-contact delta ca. 12,000 BP (Tucker 1974) and with other dated marine material in that area ranging from 11,400 to 11,950 BP (Dyck & Fyles 1963; Blake 1983). An anomalously late date of 10,200 BP (Vanderveer, pers.comm.) has been assigned to the marine incursion in southwest White Bay.

Dates from three other terrestrial sites indicate the course of ice recession in northcentral Newfoundland: a coastal site on Notre Dame Bay was deglaciated by 13,200 BP, the ice had receded inland by 11,800 BP in the Bishop Falls area and

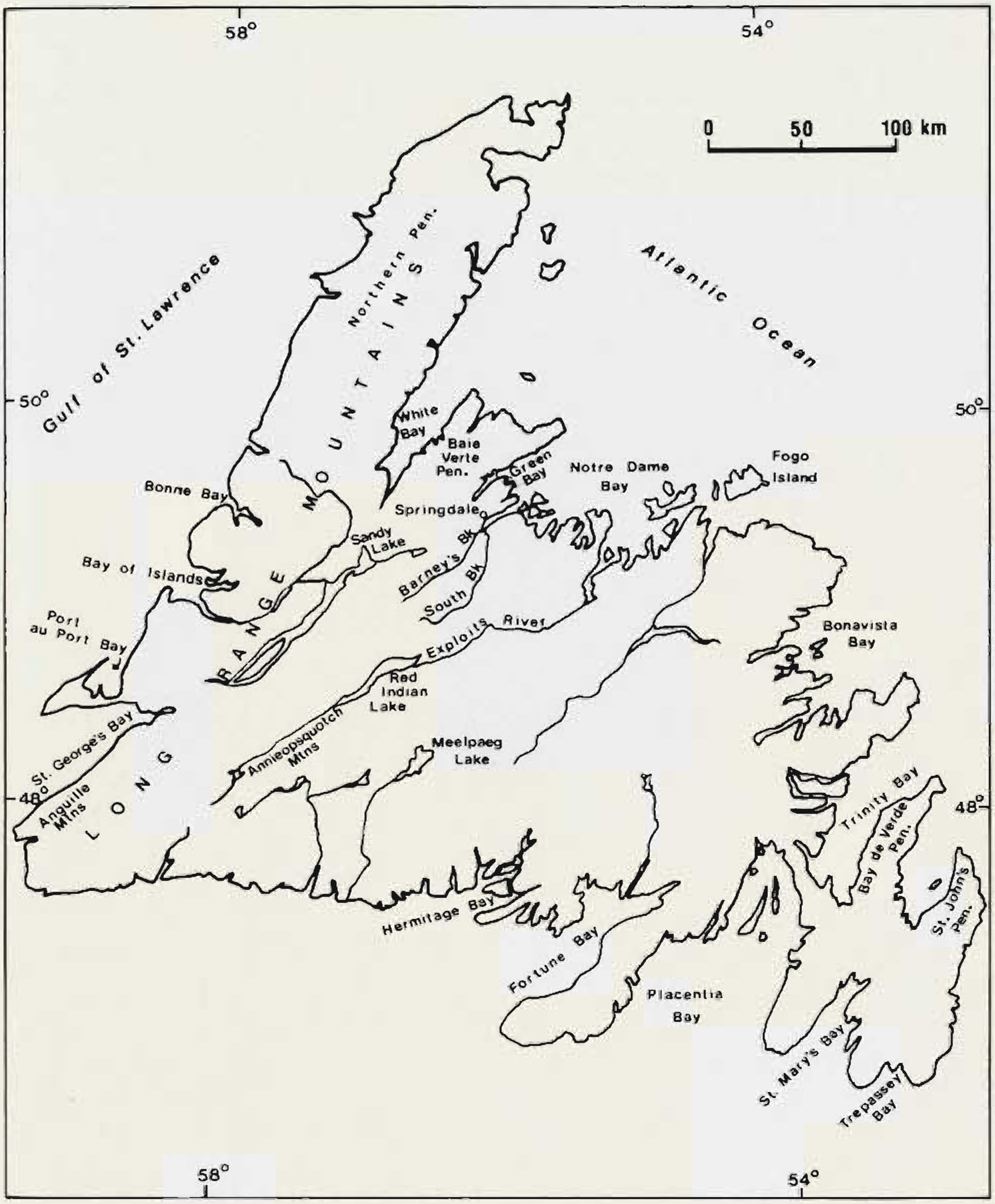


Figure 2.3: Location Map, Newfoundland.

by 11,300 BP the central plateau area was ice-free (Macpherson in Blake 1983; Macpherson & Anderson 1985). In contrast, basal lake sediment indicates that Fogo Island was not deglaciated until ca. 11,000 BP (Macpherson in prep.). In northeastern Avalon Peninsula, basal dates from ponds are younger than other coastal dates in the Atlantic region. Macpherson (1982a) demonstrated that a direct relationship between site elevation and age indicated downwasting of a final ice mass on the peninsula when the interior plateau emerged ca. 10,100 BP and sites at low elevations became ice-free ca. 9200 BP.

Throughout the Atlantic region during the last glaciation sea levels were about 100m below the present sea level (Grant 1980) so that numerous islands dotted the continental shelf. Following deglaciation of the land, and after differing periods of emergence, areas near the ice limit, including New Brunswick, Nova Scotia, Prince Edward Island and southern and eastern Newfoundland, began to submerge. Western Newfoundland, where the marine limit is between 50 and 125m above sea level, has been continually emerging until recent centuries (Grant 1980).

Raised post-glacial shorelines in the Atlantic region exhibit a general northward tilt while in Newfoundland and New Brunswick the tilt is to the northeast and northwest, respectively (Grant 1977b). This, then, accords with the model in which separate ice-caps existed in several areas in addition to the Laurentide ice sheet (Grant 1977b).

2.4 Plant Refugia

Did plants find refuge from the last glaciation on mountain summits in eastern Canada? Did ice-free coastal strips of land harbour pockets of vegetation?

Did portions of the continental shelf provide new niches for species displaced by ice sheets? Or were plants forced to migrate far beyond the ice sheets to the southeast parts of the United States? Such hypotheses have formed the basis of research for biologists and biogeographers (e.g. Fernald 1918, 1925; Wynne-Edwards 1937; Deevey 1949; Drury 1968) in northeast North America. In terms of the revegetation of Newfoundland, many of these questions have yet to be resolved.

The proposal that nunataks, or ice-free areas, provided refuge for plants during at least the last glaciation was first advanced by Fernald (1925) who identified, in parts of eastern North America, arctic-alpine species which were otherwise known to grow only in the Western Cordillera. The nunataks identified by Fernald included the Long Range of western Newfoundland, the Torngat and Kaumajet Mountains of Labrador, the Chic Choc Mountains of the Gaspé, the Magdalen Islands and Prince Edward Island. Flint (1971) added the summits of the Appalachian Mountains in New York and New England States and Grant (1977b) added Anticosti Island to the list of proposed nunataks.

As mentioned in the foregoing section, the nunatak hypothesis has found both adherents and opponents among geologists. Likewise the opinion of biologists has been divided. Wynne-Edwards (1937) opposed Fernald's nunatak theory on the grounds that the disrupted ranges of many plants could be explained by the plants' association with mainly calcareous and basic rocks.

Convincing arguments for nunatak areas come from researchers on the other side of the Atlantic. Dahl's (1946) study of bicentric distributions led him to claim that arctic-alpine flora existed along the Scandinavian coast during the last

Ice Age. Pointing to the comparatively rich extant flora on Greenland highlands, Gjaerevoll (1963) invoked the nunatak hypothesis to explain the disjunct distribution of some alpine species in Norway. He suggested that steep south-facing slopes in the Gjevilvasskammae Mountains could have offered suitable habitats for a hardy flora.

A recent study on the moss flora of western Newfoundland provides evidence for the existence of plant refugia in that part of the province. Belland (1981) identified thirty-nine moss species at Bonne Bay that exhibit a bicentric distribution, with one centre in the Western Cordillera and another in northeastern North America. Many of these mosses are listed as "forest species", but they are also found on limestone headlands, wet acidic rock faces and highland barrens. Seven of the thirty-nine species differ in their edaphic ecologies (thus opposing Wynne-Edwards' (1937) assertion) and dispersal potentials which led Belland to claim that their presence in Bonne Bay could be "best explained by their having survived glaciation at or near the sites where they occur today"(1981, p.115).

Lindroth (1963), studying the distribution of Carabid beetles in Newfoundland, espoused the view of coastal rather than highland nunatak refugia based on the inability of some of the beetles studied to endure a high-arctic climate. His conclusions point to the existence of refuge areas in western Newfoundland. For example, the occurrence of many flightless species in the southwest, where the Carabid fauna is richest, "prevents an explanation of the disjunction as due to long-range dispersal"(Lindroth 1963, p.109). In addition, he showed that in some cases the intraspecific variation of Newfoundland populations, unlike those on the mainland, tended toward subspeciation. This process, claimed Lindroth, is unlikely to have taken place in the short postglacial period.

Another refugial hypothesis concerns the possible survival of plants on the continental shelf of eastern North America and their migration to nearby land following deglaciation and eustatic rise. Terasmae (1973) proposed that New Brunswick was recolonized by vegetation from both south of the ice sheet margin in eastern North America and continental shelf refugia and that Newfoundland was revegetated primarily from the latter source. Terasmae claimed that both arctic and boreal elements grew on ice-free shelf areas extending from the Grand Banks, east of Newfoundland, south to Georges Bank, east of New York State, on the basis of peat deposits collected from Sable Island and farther south. However, in an earlier paper he attributed the abundance of coniferous pollen in pollen spectra from Sable Island ca. 7000 BP to mainland sources (Terasmae & Mott 1971). Furthermore, palynological evidence suggests that plant habitats of Sable Island have remained essentially the same over the past 11,000 BP and that the island did not support a forest community (Terasmae & Mott 1971). This may also be the case in Georges Bank where pollen analysis of fresh-water peats of ca. 11,000 BP led Emery <u>et al.</u> (1967) to propose that terrestrial vegetation covered the shelf.

Several tree species growing in Newfoundland, namely red and white pine (*Pinus resinosa* and *P. strobus*), red maple (*Acer rubrum*), dogwood (*Cornus alternifolia*), black ash (*Fraxinus nigra*) and yellow birch (*Betula lutea*) are, with the exception of the Saguenay River region, absent from the mainland area of eastern Québec-Labrador. Holland (1981) suggests that such species may have found refuge on the continental shelf during the late Wisconsin.

However, recent work compiling data from multiple sites provides strong evidence for the postglacial migration of taxa from refugia far from the ice sheet. Davis (1983), for example, has mapped the arrival of trees throughout the eastern United States by noting the radiocarbon age of the first abrupt increase in pollen influx. Isochrones were then drawn to show the position of each species' frontier at different times in the past. Her maps indicate that with the exception of white pine, which may have found refuge in the Appalachian foothills, the Coastal Plain or the continental shelf, most species migrated northeastward. In contrast to Holland's theory, most deciduous trees, including maple, first appeared southwest of the southern Appalachians. Therefore, temperate species could not have found refuge on the continental shelf.

Nevertheless, it should be noted that there was no land contact between Newfoundland and the mainland during deglaciation. In the Strait of Belle Isle, the sea level was 60m above its present position following the Ten Mile ice advance ca. 10,900 BP (Grant 1969a). In addition, the Laurentian Channel presently reaches depths in excess of 200m between Cape Breton and Newfoundland. Therefore, if sea levels during glaciation were approximately 100m below the present sea level (Grant 1980), most of this channel remained inundated. Thus had many of the modern boreal forest elements found refuge on the mainland, their only way to invade and become established on the island would have been via long-distance dispersal (e.g. by anemochore, zoochore or hydrochore modes). Certainly, many of the comparatively early dates of species' arrival in Newfoundland (see below) make the theory of postglacial migration from mainland sources alone, difficult to accept.

2.5 Sequence of Late- and Postglacial Vegetational Changes in the Atlantic Region excluding the Island of Newfoundland

During the past two decades, an increasing number of studies has been undertaken concerning the process of revegetation in eastern Canada. In the Atlantic region, investigations have been concentrated in Nova Scotia (Livingstone & Livingstone 1958; Livingstone & Este 1967; Livingstone 1968; Haddon 1975; Green 1981) although studies have also been undertaken in Prince Edward Island (Anderson 1980), New Brunswick (Mott 1975a), the Gaspé (Labelle & Richard 1981), the north shore of Québec (Mott 1976) and southeastern Labrador (Lamb 1980, 1984).

A comparison of pollen assemblage zones from Newfoundland with those of adjacent areas is presented below in the form of palynostratigraphic charts (see Figures 2.4 & 2.5). The pollen zones indicate that the development of vegetation, like deglaciation, occurred metachronously across the region.

Following the ice retreat in coastal southwest New Brunswick, a tundra-like vegetation, comprising predominantly willow and herbs indicative of open ground, colonized the area ca. 13,000 BP (Mott 1975a). Dwarf birch and aspen became more abundant by 12,600 BP but the landscape remained essentially treeless until 12,000 BP when spruce migrated into the area forming open woodlands. Spruce reached its peak 1000 years later and subsequently declined in favour of birch and aspen. Around this time a low shrub-herb tundra, similar to that recorded earlier in southwest New Brunswick, became established in Prince Edward Island and Nova Scotia (Anderson 1980).

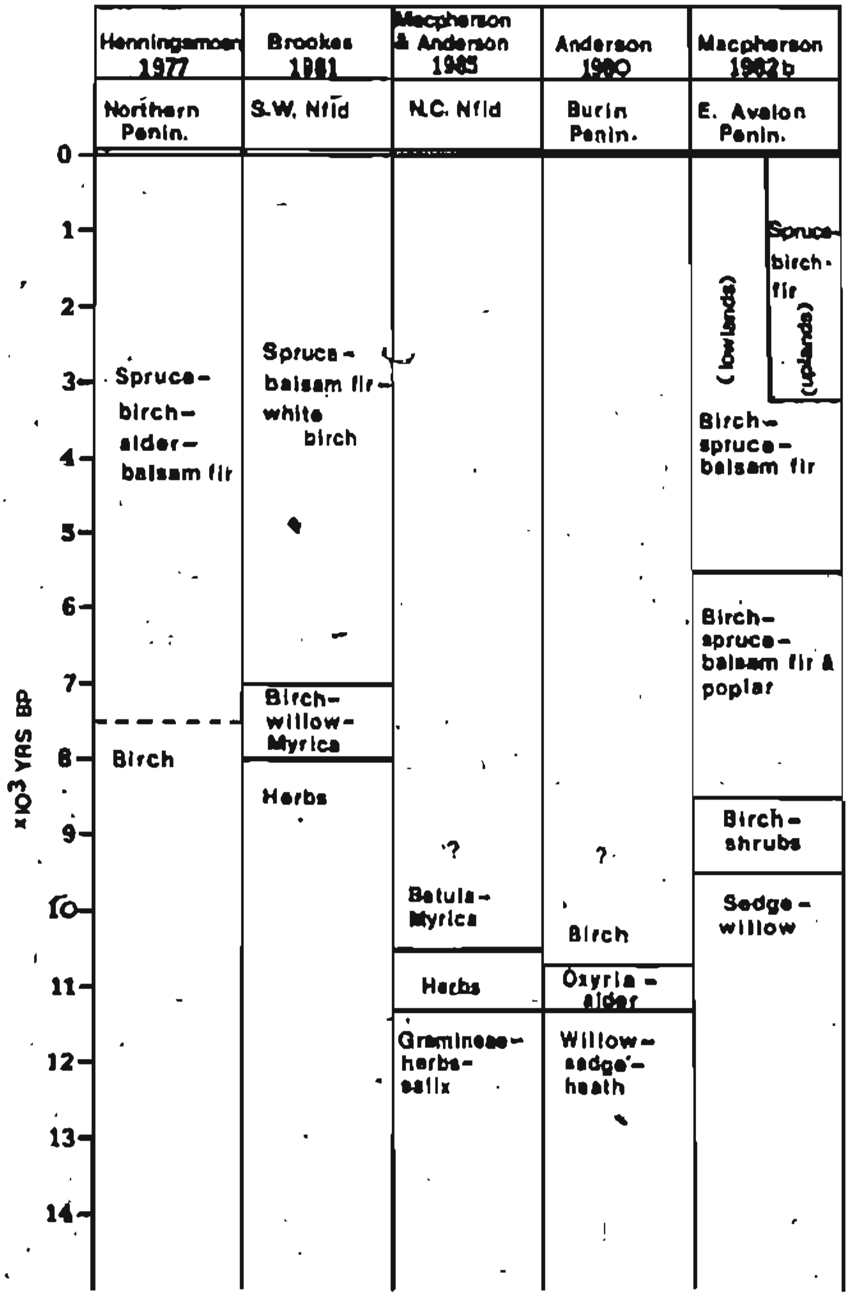
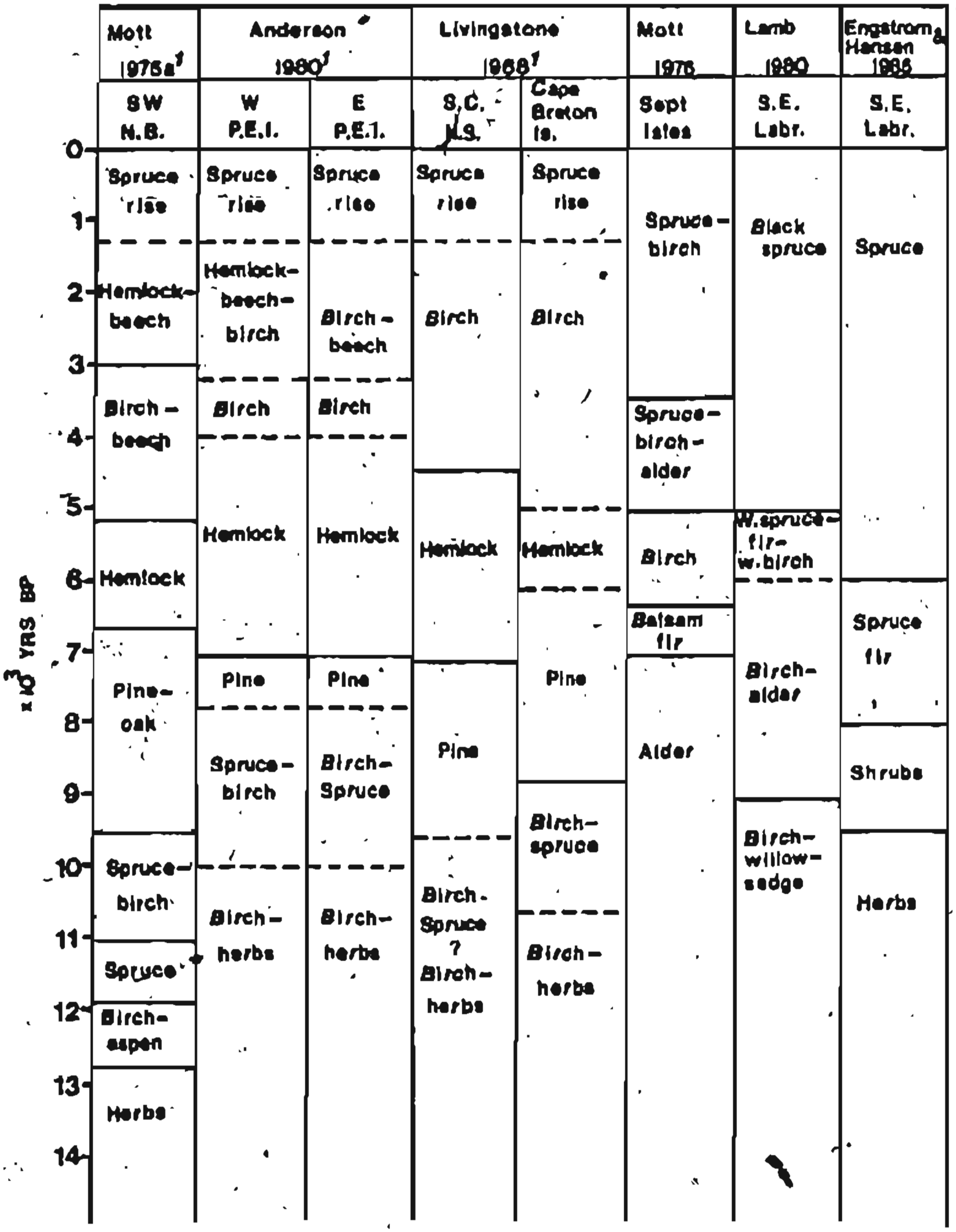


Figure 2.4: Palynostratigraphic chart, Newfoundland.



[1] from Anderson 1980, p. 1163

Figure 2.5: Palynostratigraphic chart, Atlantic Canada - excluding Newfoundland.

By 10,700 BP spruce had invaded northcentral Nova Scotia and had migrated eastward reaching the Cape Breton Island lowlands between 10,300 and 8,800 BP (Livingstone 1968), northward to the coastal plateau area of the Gaspé by 10,300 BP (Labelle & Richard 1981) and 300 years later had formed a spruce-birch association in Prince Edward Island (Anderson 1980). The arrival of spruce in the southeast of Nova Scotia and at Shaw's Bog, near Minas Basin, occurred about 1000 years later than in other parts of the province. Haddon (1975) suggested the possibility of a remnant ice cap over these latter sites that had delayed deglaciation and hindered spruce migration.

Farther north, in southeast Labrador, the landscape was sparsely covered with dwarf birch, willow and herbs by 10,500 BP (Lamb 1980). This tundra-type vegetation persisted for ca. 1500 years.

A transitional forest of dominantly pine, fir and oak developed in western Nova Scotia between ca. 9200 and 8200 BP (Green 1981), whereas in the central and eastern parts of the province, these taxa attained their maxima about one millenium later (Haddon 1975; Livingstone 1968). A birch-alder scrub had colonized the landscape of southeastern Labrador by 9000 BP (Lamb 1980). This phase lasted about 3000 years.

Vegetational development in the Maritimes between ca. 7000 and 5000 BP was more or less synchronous as hemlock migrated into the area and became the principal component of the forest (Anderson 1980). In contrast the vegetation in the northern part of the region passed through different stages of development during this period. Mott's site near Sept-Iles, Québec, (1976) had emerged and balsam fir had arrived by 7000 BP, followed by spruce and birch 600 years later.

According to Lamb (1984), a fir-dominated forest developed in southeastern Labrador ca. 6000 BP with the arrival of white spruce, balsam fir and white birch. Data from a more recent study (Engstrom & Hansen 1985) confirm this sequence but not its timing. Better dating control and macrofossil evidence allow Engstrom and Hansen to place the immigration of white spruce to southeastern Labrador at ca. 8000 BP followed by balsam fir 500 years later.

Mixed conifer-hardwood forests, in which birch and beech were the principal constituents, developed in New Brunswick and Nova Scotia after 5000 BP (Anderson 1980). The arrival of these species in Prince Edward Island was delayed several hundred years. This is attributed to the submergence of the land bridge which, prior to 5000 BP, spanned the Northumberland Strait and provided a suitable migrational pathway (Anderson 1980). Migration of hardwoods, which are less flammable than conifers, into the Maritimes, led to a decrease in fire frequency and subsequently enabled succession to proceed more rapidly (Green 1981). Until 2000 BP, the composition of hardwood forests continually changed. A resurgence of boreal species, primarily spruce, took place after ca. 2000 BP.

2.6 Palynological Research in Newfoundland

[i] Late- and Postglacial Vegetation

Comparatively few studies on vegetational history have been undertaken in Newfoundland. On the island, only two main areas, namely the Avalon Peninsula (Terasmae 1963; Macpherson 1980, 1982a,b) and the northern extremity of the Northern Peninsula (Wenner 1947; Kuc 1975; Mott 1975b; Henningsmoen 1977; McAndrews & Davis 1978) have received any detailed analysis, although research has been initiated on the southwestern and western coasts of Newfoundland

(Brookes 1981), on the Burin Peninsula (Anderson 1983), and in northcentral and central Newfoundland (Macpherson in prep.).

The pioneer work on the Island of Newfoundland was undertaken by Wenner (1947) who, in 1939, studied two ponds near St. Anthony. Wenner's pollen diagrams indicate the initial presence of a tundra vegetation followed by forest-tundra conditions and a reversion to less wooded conditions in the uppermost samples. The interpretation lacks a firm chronology, however, since the work preceded the development of radiocarbon dating.

Almost twenty years later, Terasmae (1963) published information on cores, obtained by Henderson's geological survey party, from three bog sites in the northcentral region of the Avalon Peninsula. Terasmae's pollen profiles register the migration of spruce, ca. 8400 BP, into a shrub birch and grass tundra. By ca. 7400 BP, yellow birch (*Betula lutea*) and balsam fir had arrived in the area, followed by white pine (*Pinus strobus*). The forest cover appears to have exhibited little variation after ca. 7400 BP with the exception of a reduction in fir and an increase in the pollen of 'weed' species, as a result of human influence, in the uppermost samples (Terasmae 1963).

Terasmae's work failed to provide either a precise date of deglaciation or evidence of pioneer assemblages associated with the colonization of deglaciated terrain. Recently, palynologists have favoured the analysis of lake sediments rather than peat deposits for palaeoecological reconstruction for two principal reasons. First, records from limnic deposits usually extend to an earlier stage than do those from bogs since peat development did not immediately follow deglaciation or emergence (Macpherson 1981). Second, lacustrine sediments best represent the regional vegetation whereas peat deposits reveal the local vegetation.[1] This is because lakes integrate the regional pollen rain. That is, lake sediments contain a large proportion of pollen derived from outside the basin (allochthonous pollen) via atmospheric fallout, rain-out, surface runoff and stream flow. In contrast, a large proportion of pollen in peat deposits is derived from plants growing *in situ* (autochthonous pollen)(Moore & Webb 1978). Abundant pollen producers, such as Cyperaceae, grow on peat and tend to mask the regional pollen input. In addition, in peats the local pollen component may be difficult to distinguish from the regional component because local plants may be of the same pollen type, e.g. Gramineae (Moore & Webb 1978; Jacobson & Bradshaw 1981).

Since only basal organic sediment was dated in Terasmae's study, the timing of subsequent plant species migration and vegetational changes remain speculative. An additional shortcoming of Terasmae's procedure is the width of intersample intervals, which range from 30 to 50cm in the upper portions of the cores, representing about 900 years of deposition at the Whitbourne bog site. Such coarse resolution may cause brief vegetational episodes to be missed (Short 1978). In addition, Terasmae's interpretation involved only relative (percentage) pollen counts. A comparison of relative pollen diagrams with 'absolute' diagrams from sites in northeastern Labrador-Ungava indicates that small but significant palynological changes are often masked in percentage diagrams (Short & Nichols 1977) due to the differential pollen production of plants. By contrast, 'absolute' pollen diagrams can yield more realistic reconstructions of vegetation cover.

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[1] Regional vegetation refers to large vegetation units, e.g. boreal forest, whereas local vegetation refers to minor vegetation units or plant communities.

Absolute pollen frequency, or pollen influx, is a measure of the number of grains accumulated per unit area of sediment surface area per unit of time (Birks & Birks 1980) and reflects plant population sizes. Thus, by employing absolute pollen counting techniques, the possibility of establishing changes in both the composition and density of vegetation cover is enhanced.

Subsequent palynological research on lake sediments from the Avalon Peninsula (Macpherson 1982a,b), using absolute pollen counts supported by multiple radiocarbon dates, has enabled the establishment of a sequence of vegetational changes. A sedge-willow assemblage occurred at sites of lower elevation before 9300 BP. This was succeeded by a birch-shrub vegetation lasting until ca. 8300 and ca. 7300 BP at the sites of lower and upper elevation, respectively. Spruce, followed by balsam fir, 'tree' birch and aspen arrived on the Avalon by 8,000 BP. Macpherson (1982b) discussed the dissimilarity of this phase of vegetation with the modern boreal forest and proposed that it was more like an open woodland. Profiles from Terasmae's (1963) sites indicate similar vegetational composition, although percentages of fir pollen are greater and aspen pollen was not counted.

Total pollen influx, reaching a maximum between 5300 and 3200 BP, reflects the closing of the forest and a rise in the level of the treeline in the interior plateau (Macpherson 1982b). A subsequent deterioration of the vegetation is inferred by declining pollen influx values together with a change in the stratigraphy at one of the interior plateau sites. These shifts, accompanied by a decline in the arboreal component of the pollen assemblages and relative increase in the shrub component at the other interior site, imply a retreat of the treeline in the interior plateau (Macpherson 1982b).

Other vegetational changes, recorded at one lake site in St. John's, show clearly the impact of settlement in that area: a marked decrease in arboreal pollen and a sharp increase in the pollen of 'weed' and herb species, in addition to reduced organic content of the sediment, is indicative of forest clearance and the establishment of farms (Macpherson in prep.).

Discovery of the Norse site at L'Anse aux Meadows at the tip of the Northern Peninsula provided an incentive for a number of palaeoecological investigations. Palynological research has concentrated on identifying any impact on the vegetation cover during the settlement period, ca. 1000 BP. However, results indicate that the vegetation has remained essentially unchanged since this date (Kuc 1975; Mott 1975b). Furthermore, eight pond and bog sites analyzed by Henningsmoen (1977) reveal that no significant vegetational change has taken place during at least the past 6600 years. Although a birch maximum is recorded during the lacustrine stage of one site between 7500 and 6600 BP, Henningsmoen is reluctant to comment on its possible significance primarily since changes in other pollen curves appear to reflect local plant succession rather than regional vegetational changes. Subsequent research in the area has indicated the presence of a tundra vegetation before ca. 9400 BP (Davis et al in Macpherson 1981).

In southwestern Newfoundland, Brookes' (1974) choice of his Robinson's Gully [illegible] was based on his interest in dating the Robinson's Head Moraine, and only one basal radiocarbon date (ca. 10,600 BP) was obtained, at which time a sedge-dominated tundra vegetation colonized the area (Brookes 1981). Because of poor chronological control, together with changes in the sediment type throughout the core, the sediment accumulation rate can only be estimated. Hence, only the sequence and not the timing of vegetational events can be established with any certainty. According to McAndrews (Brookes 1981), who performed the pollen analysis, it appears that the tundra phase gave way to a shrubby vegetation, principally of birch, willow and sweet gale (*Myrica gale*) which developed ca. 8000 BP followed by the invasion of spruce, balsam fir and white birch (*Betula papyrifera*) by ca. 7000 BP.

The most recent palynological studies on the Island of Newfoundland have yielded the earliest palynological records. Anderson's (1983) pollen studies on the Burin Peninsula were initiated primarily to test the hypothesis of limited Late Wisconsin ice. While the pollen record obtained is not as lengthy as was expected, the basal section of the lake core has provided the earliest Late Wisconsin terrestrial radiocarbon date and pollen assemblages for the Island. Basal gyttja records a sedge-shrub tundra ca. 13,400 BP. A subsequent transition to a herb-tundra by 11,300 BP was followed by a reversion to shrub-tundra conditions ca. 10,700 BP.

A similar tripartite sequence of vegetational change has been identified at a site near the coast of Notre Dame Bay. The initial vegetation following deglaciation is interpreted (Macpherson & Anderson 1985) as a sparse herb-dwarf shrub tundra, with a floral composition not unlike that of the present-day vegetation of northern Greenland. By 12,400 BP, a reversion to a sparse herb-tundra is recorded. This phase, lasting until ca. 10,500 BP, was replaced by a tall-shrub tundra and is interpreted as being similar to the low-arctic dwarf-shrub heath of coastal south Greenland (Macpherson & Anderson 1985). Spruce arrived at 9600 BP (GSC-4183; Macpherson in prep.) followed soon after by balsam fir.

An increase in pine concentrations, possibly indicating the beginning of the hypsithermal, occurred at 5960 BP (GSC-4086; Macpherson in prep.), and at 6340 BP (GSC-4148; Macpherson in prep.) at a site 65km to the south. The subsequent decline of pine occurred at 4200 BP (GSC-4107; Macpherson in prep.) at the coast..

Analysis of lake sites in the Bishops Falls and central plateau area is still in progress. In the Bishops Falls area, a low-shrub tundra-type vegetation followed deglaciation ca. 11,800 BP. Pollen influx values are similar to those from contemporary tundra sites in coastal west Greenland (Macpherson in prep.). Spruce arrived at an unprecedentedly early date of 10,800 BP (GSC-4131; Macpherson in prep.), and pollen concentration values for balsam fir suggest that this species migrated to the area about 300 years later. Farther south, on the central plateau, the earliest recorded vegetation ca. 11,300 BP is shrub-tundra, with successive peaks of willow, sweet gale, birch (probably shrub) and juniper. A second rise in birch (probably tree) follows these events and a spruce rise occurred at 8530 BP (GSC-4122; Macpherson in prep.).

[ii] Late- and Postglacial Climate

Lithologic and pollen evidence from coastal sites in south and northcentral Newfoundland indicate a late-glacial climatic oscillation between 11,300 and ca. 10,700 BP (Anderson 1983; Macpherson & Anderson 1985). This evidence corroborates similar findings in the Maritime provinces where analysis of peat and lake sediments at over twenty sites has revealed a climatic reversal between 11,000 and 10,000 BP (Mott 1983; Mott et al. 1984).

Pollen analysis of the two Newfoundland sites indicates that the initial period of postglacial warming was interrupted ca. 12,400 and 11,300 BP at the north and south sites, respectively (Anderson 1983; Macpherson & Anderson 1985). Distinct changes in both the composition and abundance of vegetation, recorded in the north until 10,500 BP and in the south until 10,700 BP, are attributed to cooler summers with longer-lasting snowbeds (Macpherson & Anderson 1985). The resumption of postglacial warming is evidenced by the near disappearance of tundra species, the increase in total pollen influx and the change from mineral to more organic sediment (Macpherson & Anderson 1985).

The greater abundance of vegetation at the south coastal site during the late-glacial is ascribed to the site's proximity to comparatively warm waters (Macpherson & Anderson 1985): foraminiferal evidence points to the warming of the ocean south of the Grand Banks ca. 13,000 BP (Ruddiman & McIntyre 1981). The southern site was exposed to warm, moist southerly winds whereas the northern coast was swept by winds cooled by the island's residual ice cap (Macpherson & Anderson 1985). A comparison of pollen spectra from the two sites led Macpherson & Anderson (1985) to claim that the reversion of vegetation was not a function of changing palaeowind direction, but rather due to a cooling of the ocean surface south of the island.

These pollen studies by Anderson (1983) and Macpherson and Anderson (1985) are concerned with late-glacial (i.e. end of the Late Wisconsin) climatic oscillations. Previous palynological work in the Atlantic region has focussed primarily on postglacial or Holocene climatic changes.

Evidence of a climatic optimum and subsequent deterioration during the Holocene has been most apparent in records from the Maritime provinces.

Increases in pine and oak pollen between ca. 9000 and ca. 7000 BP in the Maritimes are ascribed to warmer temperatures (Livingstone 1968; Mott 1975a; Anderson 1980).

An expected result of a warming climate would be the northward range expansion of southern species (Lamb 1980). However, neither pine nor oak extended their range to southern Labrador and white pine appears to have arrived late in the Avalon Peninsula (Terasmae 1963). Evidence of climatic amelioration in these two areas is difficult to disentangle from the effects of other processes (e.g. migrational lag). Maximum pollen influx values suggest only a continuing amelioration of the climate until ca. 4000 BP in southern Labrador (Lamb 1980). On the Avalon Peninsula, different indicators in the pollen record point to a series of stages of postglacial warming (Macpherson 1982b). A rapid increase in spruce and balsam fir pollen concentration suggests a warming of the climate between 9300 and 8300 BP. From the continuous presence of aspen, it appears that conditions were warmer and drier from 8300 to 5400 BP than at present. The period of maximum warmth and moisture occurred from 5300 to 3200 BP when arboreal pollen influx and sediment accumulation rate reached peak values.

Lamb (1980) attributes a reduction in lacustrine organic productivity and diminished pollen influx values after 2500 BP in southern Labrador to a cooling trend. Subsequent palynological and geochemical analyses of lake sediments in nearby areas point to a climatic deterioration after 4000 BP (Engstrom & Hansen 1985). In the Maritime provinces, a broadly synchronous reversion of the vegetation after ca. 2000 BP, indicated by an increase in boreal trees and a decline in hemlock and hardwoods, notably birch and beech, is generally ascribed to climatic deterioration (Mott 1975a; Anderson 1980; Green 1981). In southeastern Newfoundland climatic cooling has been invoked to explain reductions in pollen influx and sediment accumulation and a retreat of the treeline in the interior plateau after ca. 3000 BP (Macpherson 1982b).

CHAPTER 3 PURPOSE OF PRESENT STUDY

The foregoing review of palynological studies in insular Newfoundland points out the fragmentary nature of available evidence pertaining to (i) Late Wisconsin ice limits and pattern of glacial recession, (ii) the late- and postglacial vegetational history, and (iii) the late- and postglacial climate of Newfoundland.

With respect to the postglacial development of vegetation in Newfoundland, only the northeast Avalon Peninsula has been treated in detail. Other studies lack adequate dating control of vegetational events (e.g. Brookes 1981); lack a comprehensive evaluation of pollen data due to (i) the use of percentage data only, or (ii) wide intersampling intervals (e.g. Terasmae 1963; Henningsmoen 1977), or (iii) the absence of differentiating important taxa such as *Picea*, *Betula* and *Pinus* species (e.g. Henningsmoen); have concentrated on early plant recolonization (e.g. Anderson 1983; Macpherson & Anderson 1985); or have focussed on the vegetational history of only the last 1000 years (e.g. Kuc 1975; Mott 1975b).

The present study was undertaken, therefore, to add to the knowledge of Late Quaternary palaeoenvironments of the province. It was proposed that the research involve the analysis of lake cores from the Baie Verte Peninsula, northcentral Newfoundland. Pollen analysis and loss-on-ignition analysis would be employed to investigate the vegetational history of the area, and radiocarbon dating would be used to provide a minimum age for deglaciation in addition to dating specific vegetational events. The Baie Verte Peninsula was chosen primarily because of the absence of information on the vegetational history for the area, and the study would thus provide a link between studies in the east, north and west parts of the province, and secondarily to yield possible evidence relating the the limit of Late Wisconsin ice.

In the planning stages of the research, a decision to sample one lake only was made. However, during the field expedition, a complete sequence of deposits was retrieved from two lakes. After preliminary pollen analysis it was decided that one lake site be used to investigate the entire Holocene pollen record and the other be studied to provide additional data on plant recolonization during the early Holocene.

3.1 Objectives

(i) The primary objective of this study was to investigate the Late Quaternary vegetational history of the northern Baie Verte Peninsula. This involved tracing the nature of changes that occurred in the vegetation from the initial colonization through to the present-day formation.

(ii) Ancillary objectives were:

Vegetation:

(1) to determine the arrival of the main arboreal species using pollen concentration values and radiocarbon dates; and to compare the arrival of these species with their arrival at other sites, from previous studies, in an attempt to establish migration routes,

(2) to ascertain the order in which species of *Picea* and *Betula* (in terms of 'tree' and 'shrub') arrived in the area,

(3) to discover whether the range of *Pinus resinosa* was more extensive in the past, i.e. whether the species' present distribution (see Figure 3.1) represents a relict distribution.

Climate:

(1) to attempt, by interpretation of postglacial vegetational changes, to determine the sequence of any climatic changes, and place the climatic history of the area within the regional (Atlantic) framework,

(2) to investigate the early stratigraphic sequences and pollen trends in order to identify any late-glacial climatic oscillation similar to that recognized by Anderson (1983) and Macpherson and Anderson (1985).

Deglaciation:

(1) to analyze and radiometrically date pollen assemblages from the basal core segments in an attempt to determine the timing of deglaciation. To reach this objective the following criteria had to be met:

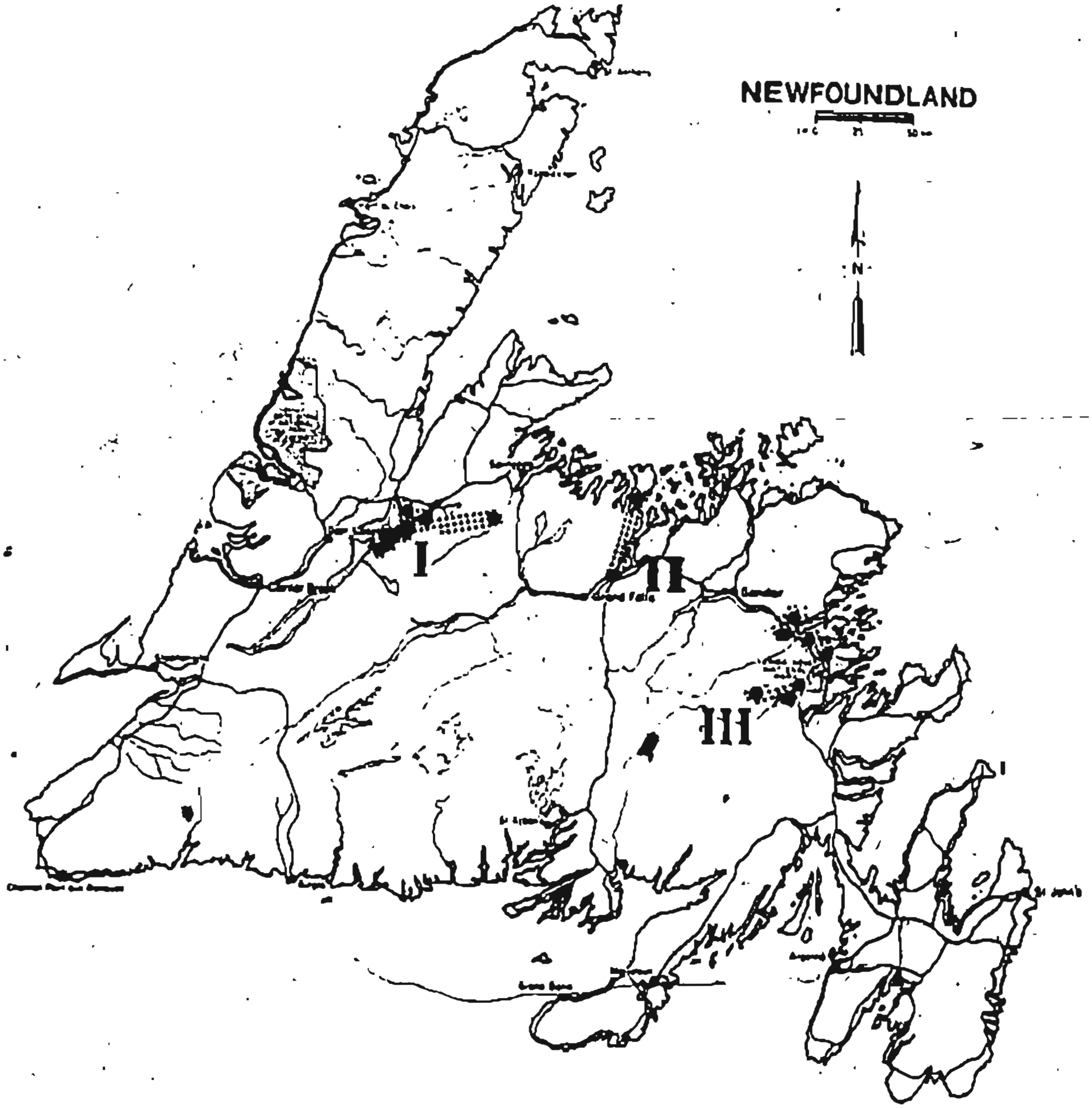


Figure 3.1: Present distribution of *Pinus resinosa* Ait., showing the geographical area of 20 stands in three lines of occurrence.
(Source: Roberts 1985)

(a) the sites are above the marine limit
(b) the basal dates are early for northcentral Newfoundland
(c) the pollen assemblages of the dated material indicate an early stage in the development of the vegetation (Macpherson 1982a).

Late Wisconsin Ice Extent:

(1) Alternatively, it was proposed that were the lake cores to provide basal radiocarbon dates significantly earlier than those of the north-central coastal terrestrial site (Macpherson & Anderson 1985) and the Halls Bay Delta site (Tucker 1974), it would be possible to confirm the hypothesis of limited ice in the area as proposed by Grant (1977b) and Prest (1984).

CHAPTER 4 METHODOLOGY

4.1 Field Work

[i] Site Selection

Several potential lake sites on the Baie Verte Peninsula were selected from 1:50,000 topographic maps and colour air photographs based on their accordance with the following criteria:

(1) above the marine limit, i.e. above 75m (Grant 1980).

(2) within 200m of a road, for ease of access.

(3) no through current, to avoid sediment flushing and redeposition.

(4) with littoral sediment, as those without are likely to have "strong, temporally variable patterns of sediment focussing" (Davis et al. 1984, p.288).

(5) preferably small, and with a regular basin morphology, to minimize the effects of sediment focussing.

(6) an area free of carbonate-rich bedrock, to avoid contamination in the radiocarbon assay.

During field investigations in June 1984, two lakes (Compass Pond, 50°02'03"N 56°11'47"W and Small Scrape Pond, 49°56'59"N 56°05'17"W) were chosen as sampling sites. The remainder were eliminated owing to difficulty in access, possible disturbance of sediments (due to recent road or powerline construction near the lake margin), or water depths beyond the range of the coring equipment. The deepest parts of the sampled lakes were located with a plumb line. At both sites, however, the greatest water depth exceeded the total length of the coring equipment, so sampling sites were located near the lake centres within the range of the equipment.

[ii] Coring

The lakes were sampled from two anchored boats bolted together with a platform through which coring operations were performed. A modified Livingstone piston corer, retrieving core sections with a 5cm diameter and 1m length, was used. A heavy plastic casing, extending from the platform to the lake floor, ensured the vertical position of the extension rods and coring tubes down through the sediment.

Duplicate cores were retrieved from each site; one to be used for loss on ignition, the other for pollen analysis and radiocarbon dating. To ensure a high degree of correlation between the cores of each site, the second core was obtained from within a metre of the first. Seven sections were collected for each of the Compass Pond (CP) cores. At Small Scrape Pond (SSP), the first core had four sections, and the second core had three sections.

The cores were extracted in the field, to ensure that complete sequences had been obtained, and placed in clean, rigid plastic troughs. The troughs were labelled, then plugged, sealed with plastic wrap and aluminum foil and transported horizontally to the laboratory for analysis. Compression of core segments upon extraction was minimal (averaging 3-4cm) and was adjusted for in the sampling procedures.

[iii] Vegetation Survey

A survey of the modern flora and vegetation types in the surrounding landscape was made as an aid to palaeoecological interpretation. A reconnaissance survey took place during the initial coring expedition. In August, 1984 and August 1985, the sites were revisited and a more detailed list of the flora immediately surrounding the lakes and within the drainage basins was compiled. Most of the aquatic plants, and taxa occurring in the herbaceous, shrub and canopy layers were identified to the species level either in the field or in the university herbarium.

4.2 Laboratory Analysis

[i] Stratigraphy

In the laboratory, the surface of the cores were scraped clean to avoid contamination as a result of smearing during core extraction. Description of the sediments included colour, using a Munsell Soils Colour Chart, composition (gyttja, clay, sand), depth at which composition changed, and general consistency (fibrous, firm, jelly).

[ii] Radiocarbon dating

Core segments with a vertical thickness of 5cm were extracted at intervals along the cores and dispatched to the Geological Survey of Canada for radiocarbon dating. Since the initial object of the study was to analyze one site, it was decided that only the CPII core required several radiocarbon dates. Sampling was guided by (a) changes in sediment composition, and where these did not occur, by (b) regular intervals. One sample was taken from the basal clay-gyttja (540-545cm) in order to determine the minimum date of deglaciation. A sample from the transition from gyttja to clay-gyttja (435-440cm) was extracted in order to calculate the sediment accumulation rate of the clay-gyttja layer. Five other sections were extracted at regular intervals in the gyttja permitting calculation of sedimentation rates and pollen influx.

Of the SSP cores, only one exhibited a complete sequence of deposits. The basal 5cm of organic material (282-287cm) of SSP II was radiocarbon dated to determine the timing of deglaciation of this site and provide a better control on the basal date for the CP cores. The basal 5cm (315-320cm) of SSPI was also radiocarbon dated.

(iii) Loss on Ignition

Loss on ignition analysis was performed on samples of the duplicate cores (CPI and SSPI) in the fall of 1984. The ignition loss provides an approximation of the organic matter content of the sediment (Bengtsson 1979). Changes in the percentage of organic matter content throughout a core is frequently related to terrestial biomass productivity (cf. Lamb 1980).

Samples of 1cm vertical thickness were extracted every 5cm from CPI. The intersample interval for SSPI ranged from 4cm to 27.5cm. The procedure used (after Bengtsson 1979) is as follows:

(1) A porcelain crucible was ignited for one hour at 550°C in a furnace (Sybron/Thermolyne 1400 Electric type), cooled to room temperature in a desiccator and the <u>weight of the crucible</u>(a) was determined.

(2) A sediment sample was transferred to the crucible and dried overnight in an oven at approximately 105°C.

(3) The sample was cooled in a desiccator and the <u>weight of the crucible and oven-dried sample</u>(b) was determined. The sample was pulverized in a mortar, to ensure its complete ignition in the next step, and

returned to its crucible.

(4) The crucible with the dried sample was placed in a preheated furnace at 500-550°C for approximately 2 hours. Lids were placed on all crucibles to avoid ash losses. Samples not fully ignited were removed, stirred with a metal wire and returned to the furnace.

(5) After cooling in a desiccator, the weight of the crucible and ignited sample(c) was determined.

The calculation for loss on ignition was:

$$\frac{b - c}{b - a} \quad X \quad 100\ \%$$

[iv] Macrofossils

Although the cores were not systematically analyzed for macrofossils, a few plant remains were discovered and identified to at least the genus level.

4.3 Pollen Analytical Methods

[i] Sampling and Processing

Subsamples of the CPII core were extracted at 5cm intervals. Pollen analysis of this core, however, was undertaken first at 10cm intervals to see if dissimilarities between adjacent samples warranted closer sampling. Closer sampling was performed throughout most of the clay-gyttja layer, at the base of the gyttja layer, and at several levels at which certain taxa peaked. It was decided that analysis of the SSPII core would provide an interesting comparison of early vegetation stages and migration of important taxa. Therefore, only the basal portion of this core (from 180 - 287cm) was sampled, at intervals of 5-10cm.

The standard method used for concentrating pollen in sediments (Faegri & Iversen 1975) is outlined below.

(1) 10% KOH (potassium hydroxide) treatment to deflocculate and remove humic acids - KOH was added to 1ml sediment samples which were then heated to boiling in a water bath, strained through a Gooch crucible to remove coarse organic debris, and rinsed with distilled water.

(2) Acetolysis to digest cellulose - samples were first washed in glacial acetic acid. This dehydrates the material since the next step is exothermic. Material was acetolysed in a mixture of acetic anhydride (CH_3CO_2O) and concentrated sulphuric acid (H_2SO_4) and boiled in a water bath for 5 minutes, then washed with glacial acetic acid (to remove excess H_2SO_4) and given several rinses with distilled water.

(3) Staining and dehydration - safranin was added to the polleniferous residue and the material was dehydrated with tertiary butyl alcohol. Staining facilitates the distinction between pollen grains and other material and increases the contrast of morphological details.

(4) Mounting - the material was mounted in silicone oil. This mounting medium has several advantages including

a low refractive index proving good contrast for visual observation; the possibility of rotating grains for easier identification, and does not appear to cause swelling (Andersen 1965).

Samples that had a high minerogenic component were processed with hydrofluoric acid. The treatment, inserted after the KOH procedure and prior to acetolysis, involved boiling the sample first in HF (~4.8 % solution) and then in dilute HCl in a water bath, each for 5 minutes, to remove siliceous material.

For clay-rich samples, both chemical and mechanical treatments were employed. The former, inserted prior to the KOH process, consisted of washing the sample several times with a 5% sodium pyrophosphate ($Na_4P_2O_7$) solution, and decanting the liquid between two beakers during the last wash to remove sand particles. The latter treatment, inserted after acetolysis, entailed resuspending the sample in $Na_2P_2O_7$ and filtering clay particles through a 7 μm mesh screen (Cwynar et al. 1979).

[ii] Absolute Pollen Extraction Technique

The foregoing section allows for only the 'relative' frequency, or the percentage of each pollen type to be determined. To permit the calculation of 'absolute' pollen frequencies, so that abundances of fossil pollen types are independent of each other, an exotic marker grain method was employed. A measured volume of a suspension of exotic pollen (*Eucalyptus*) of known concentration was added to the 1ml sediment samples prior to processing enabling the determination of fossil pollen concentration ($grains \cdot cm^{-3}$) as follows:

$$\frac{\text{Fossil pollen counts}}{\text{Exotic pollen counts}} \times \text{added pollen concentration}$$

and, together with sediment accumulation rates, as determined by successive radiocarbon dates, the annual pollen deposition rate, or pollen influx (grains·cm^{-2}·yr^{-1}), was calculated as

Fossil Pollen Concentration X Sedimentation Rate

[iii] Microscopy

Pollen identification and counting was performed over a seven month period. A Carl Zeiss Jena Laboval 2 binocular microscope was used at a total magnification of x800, using a x40 objective, with x12.5 oculars and a binocular inclined tube factor of x1.6. The initial counting of grains at this magnification in samples from the clay layer of CPII proved to be time consuming even though the sediment volume to exotic suspension ratio was adjusted to take into account low pollen concentrations (500-3300 grns·cm^{-3}). Since these samples contained little extraneous material to obscure pollen grains, the levels were subsequently scanned at a magnification of x200, using a x10 objective. Examinations of rare and difficult grains were made at a magnification of x2000, using a x100 oil-immersion objective. Grain measurements were determined using an ocular micrometer scale, with one division equalling 1.5μm at a total magnification of x800. Traverses were made at regular intervals of 2mm across the microscope slides to avoid bias caused by the differential movement of grains of various sizes.

[iv] Pollen Sum

Pollen analysis involves a sequence of sampling. A sample of the lake sediment is taken by coring; samples within the core are selected for processing;

next, samples of the polleniferous residue are mounted on microscope slides. Finally, the palynologist must decide the number of pollen grains to count to provide the representative proportions of grains settling on and incorporated into the lake sediment.

The selection of taxa included in the pollen sum is based on the members of the population under study. Since the purpose of this research was directed not exclusively at the forest history but includes the sequence of vegetational changes following colonization, the pollen sum chosen comprised shrub and herb (including Cyperaceae) as well as tree pollen. This group of taxa was regarded as representative of the regional pollen rain. The percentage frequency of any taxon, e.g. *Picea*, within this land pollen sum (ΣLP) was calculated as:

$$\frac{\Sigma \textit{Picea}}{\Sigma \text{LP}} \times 100$$

The percentage frequency of any other group namely spores, aquatics, *Pediastrum*, indeterminate, and unknown, was calculated by adding the sum of the land pollen to the sum of that group, for example, ΣLP + ΣAquatics, thereby precluding the possibility of any taxon reaching over 100% (Birks & Birks 1980).

The number of grains counted is dependent upon the number of taxa in a sample and enough grains should be counted to maintain constant percentages of the pollen sum (Birks & Birks 1980). A minimum count of 300 grains, based on previous studies in Newfoundland (e.g. Macpherson & Anderson 1985), was judged as a reasonable pollen sum for this study. However, in order to attain this sum for the basal minerogenic samples, it was calculated that over 30 slides per sample would require scanning. It was therefore decided that counting extra levels rather than additional grains per sample would be more profitable.

In the gyttja layer of CPII (0-440cm), the pollen sum of 300 was exceeded by an average of 48 grains, requiring the scanning of approximately 3 slides per level. In the upper portion of the clay-gyttja transition (440-505cm) the pollen sum averaged 311 grains, and approximately 4 slides per level were scanned. In the lower portion of the clay-gyttja (505-545cm), pollen sums averaged 111 grains, requiring approximately 6 slides per level. In the clay layer (545-620cm), pollen sums averaged 40 grains, requiring approximately 6 slides per level.

[v] Pollen Diagram Construction

(1) Pollen diagram format

The main results have been presented in the form of pollen percentage, concentration and influx diagrams (Chapter 7 & Appendix). A similar format was used for all the diagrams to facilitate their comparison although not all the taxa appearing in the percentage and concentration diagrams are included in the influx diagram. The POLSTA computer-graphics program was employed for data manipulation and diagram construction: although pollen percentages and concentrations were determined prior to obtaining the computer package, radiocarbon dates and sample levels were read into the program, enabling it to calculate sedimentation rates and influx values.

On the far left of the diagrams (see Figures A.1, A.2, A.3, A.4, A.5) are shown the radiocarbon dates (in corrected radiocarbon years BP) and the standard deviations. To the right, the sediment lithology is represented schematically, with descriptions at the base of the diagram. Adjoining the lithology column are depths of the core segments. The depth (in cm) below surface from which samples were taken for pollen analysis is shown to the right of the core segments. Individual curves for pollen and spore types identified are given next and drawn to a standard scale. The occurrence of rare elements (e.g. *Epilobium*) is designated by a letter, the code for which is given at the base of the diagram. This main part of the diagram is divided into physiognomic categories: trees, shrubs, herbs, pteridophytes, aquatics, alga, with indeterminate and unknown forming separate categories. The categories are organized in such a way that those taxa occurring most frequently throughout the profile are grouped at the left. The exception is *Betula* which, because its pollen may be of 'tree' or 'shrub' origin, is placed on the far left of the individual pollen curves. Further ordering of taxa within categories is alphabetic with related genera grouped by family.

To the right of these curves on the pollen concentration diagrams are two columns of figures: the first is the pollen sum (ΣLP) and the second is the total number of pollen grains and spore types identified (Σtaxa).

On the far right of the diagrams are shown the results of the numerical zoning program, CONSLINK, together with the local pollen zones: the zonation procedures are described in the following section.

All three types of diagrams, viz. percentage, concentration and influx, are presented in this study because a different set of information can be elicited from each of them. Pollen percentage diagrams for example, indicate the taxa present and composition of plant communities yet, by their very nature, do not register changes in plant abundance. Pollen influx diagrams, on the other hand, can provide information on the absolute abundance of individual taxa and vegetation density (Davis & Webb 1975). But pollen influx diagrams are also subject to misinterpretation if sedimentation processes are overlooked. For example, sediment focussing may cause misleading changes in pollen influx without parallel changes in pollen concentration. It is primarily for these reasons that all three diagrams (at least for one lake site) are presented. The construction of other diagrams (e.g. age/depth curve, loss-on-ignition) are described and explained in the appropriate chapters.

(2) Pollen zonation

To facilitate the description, comparison and interpretation of pollen sequences, the pollen diagrams have been divided into pollen zones (and subzones) here defined as "a body of sediment with a consistent and homogeneous fossil pollen and spore content that is distinguished from adjacent sediment bodies by differences in the kind and frequencies of its contained fossil pollen grains and spores" (Birks 1973, p.273). These zones are, more accurately, pollen assemblage zones in that they are "named from one or more taxa particularly prominent or diagnostic of the assemblage, although name-givers need not be confined to the zone or found in every part of it" (Amer.Comm. on Stratigraphic Nomenclature 1961, Article 21(b), p.656). Pollen zones have been delimited for each of the two lake sites and are, as such, site specific and referred to as local zones. Certain similarities between the sites have been identified, enabling a regional pollen zonation to be proposed for the study area.

Pollen zone boundaries were located initially by visual inspection of the percentage diagrams and were subsequently confirmed, for the most part, by results of a numerical zonation procedure, CONSLINK (after Birks 1979; revised by Maher n.d. and pers.comm.). CONSLINK is a constrained single-link analysis and the results represent the values at which the levels are joined together in a dendrogram: the lower the value, the more similar are the levels (Birks 1979). Maher's revision of the program differs in that two routines have been produced: the first, CONSLINK MIN, is based on that of Birks and ties on the most similar item after the first; the second routine, CONSLINK MAX, is perhaps a more rigorous test of similarity in that the least similar item is tied on after the first.

With this program, it is standard procedure to include (a) only pollen and spores of non-aquatic taxa, and (b) only those pollen and spores whose values attain at least 5% of the basic pollen sum in at least one level in the profile (Birks 1979). Up to eighteen pollen types may be handled: with the Compass Pond II data, these included *Betula*, *Picea*, *Abies*, *Pinus* (total), *Alnus*, Coryloid, *Myrica*, Ericales, *Juniperus*, *Salix*, Gramineae, Cyperaceae, *Ambrosia*, *Artemisia*, Compositae, *Sphagnum*, *Lycopodium* (total) and Filicales; with the Small Scrape Pond data, the following twelve taxa were considered: *Betula*, *Picea*, *Pinus* (total), *Alnus*, Coryloid, *Myrica*, *Juniperus*, *Salix*, Cyperaceae, *Artemisia*, *Lycopodium* (total) and Filicales.

[vi] Pollen and Spore Identification

Interpretation of the fossil pollen record is enhanced with the identification of pollen grains to the species level. In Newfoundland, this is particularly true of the genera *Betula*, *Picea* and, to a lesser extent *Pinus*, the species of which are indicative of particular ecosystems (e.g. forest-tundra or boreal forest) and ecological conditions (e.g. acid or alkaline soils; wetland or upland sites). In many instances, however, the morphological similarity of members within a genus, and indeed within a family in the case of many herbaceous taxa, often precludes identifications to a lower taxonomic level.

In this investigation, pollen and spore identification was primarily to the genus level. The determination of rare and difficult grains was facilitated by reference to several keys (primarily McAndrews et al. 1973; Faegri & Iversen 1975; Moore & Webb 1978; Andrew 1980) and a reference collection of over 200 modern samples. Certain taxonomic determinations are described below:

Betula

Several 'tree' and 'shrub' species of the genus *Betula* occur in Newfoundland. Morphological evaluation of *Betula* grains is not a suitable method for determining their parent source. However, previous studies have demonstrated the possibility of using grain diameter and grain diameter:pore depth ratio in differentiating groups of species (Leopold 1956; Clausen 1962; Birks 1968). A recent study of five Newfoundland species of *Betula* (Dyer 1981) indicated that it is possible, on the basis of size-frequency curves, to distinguish 'tree' birch from 'shrub' birch species although an overlap exists. The diameter of *B. papyrifera* and *B. lutea* grains ranges from 24.5 to 35μm, whereas the diameter of *B. pumila* and *B. michauxii* grains ranges from 18.5 to 26μm. In addition, the diameter of *B. cordifolia* extends from 20 to 34 μm. Following that study, thirty fossil birch grains were measured from 12 levels from CPII and from 3 levels from SSPII. Grains below 20μm are classified as 'shrub' birch. Because of the overlapping size ranges, grains in the range of 20 to 26μm may have come from either arboreal or shrub sources.

Picea

Both *Picea mariana* and *P. glauca* occur in Newfoundland. In the study by Dyer (1981), the application of the discriminant analysis method of Birks and

Peglar (1980) to modern spruce pollen from Newfoundland demonstrated the possibility of distinguishing these species (see Figure 4.1). The method involves three stages of identification: first, the discrimination of modern reference material of known taxa is made; second, the fossil grains are assigned to one of the reference groups; and third, the fossil grain is appropriately named (Birks & Peglar 1980). In the present study, the method of Birks and Peglar, employing a modern pollen data set from samples collected in Newfoundland, was applied to twenty fossil spruce grains from each of six levels of CPII. Six quantitative variables were measured for each grain (total grain length, corpus breadth, corpus height, saccus width at base, saccus height and maximum saccus breadth) and a linear discriminant analysis was applied to the data. By measuring these variables, the position of each fossil grain on the discriminant function (see Figure 7.5) was determined as:

$$Rf = x_1\lambda_1 + x_2\lambda_2 \ldots\ldots x_6\lambda_6$$

where,

Rf is the position of the fossil grain on the discriminant function.
$x_1 \ldots x_6$ are the variables.
$\lambda_1 \ldots \lambda_6$ are the discriminant function coefficients as given in Birks & Peglar (1980).

During the analysis, however, it was noticed that, because only whole grains positioned in equatorial view were used, the method may involve a morphometric bias. That is, smaller grains may be less likely to crumple and are easier to rotate than are larger grains.

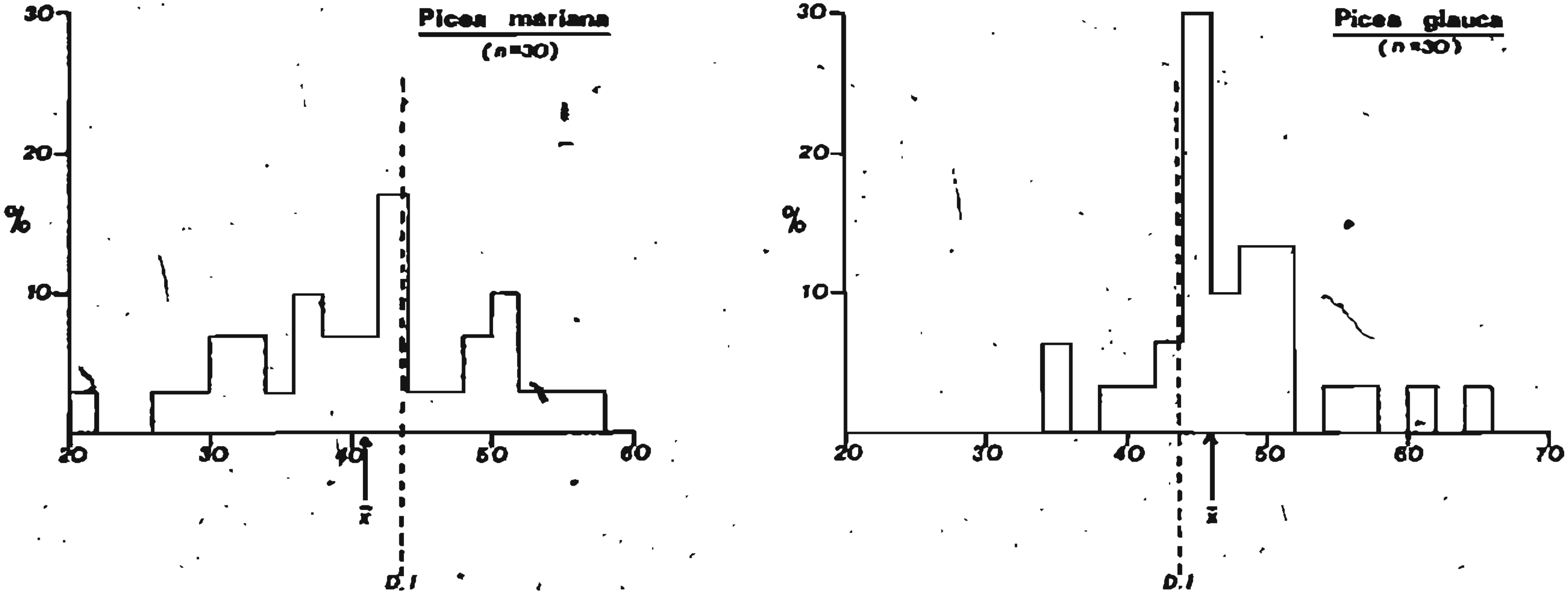


Figure 4.1: Positions of modern pollen of Newfoundland species of Picea mariana and P. glauca on the linear discriminant function based on six quantitative variables.

(Source: Dyer 1981)

Pinus

Two indigenous species of *Pinus* occur in insular Newfoundland. *P. strobus* occurs throughout most of the island except the Northern Peninsula, but *P. resinosa* is limited in its range to central Newfoundland (*sensu* Damman 1983). A preliminary examination of modern pollen of these species indicated the possibility of their differentiation. In *P. strobus*, nodules occur on the surface of the grain between the bladders whereas these features are generally absent in grains of *P. resinosa*. Positive identification is not always possible due to deterioration (particularly in the basal samples), concealment (by folded bladders or extraneous matter) or inability to rotate grains to the required position. These grains were entered as *Pinus* undifferentiated.

Alnus

Both *Alnus rugosa* and *A. crispa* grow in the study region. However, no morphological characteristic nor size criteria have been found which adequately distinguish the species (Dyer 1981), and the two were not differentiated in this study.

Coryloid

This category includes any triporate grains that due to deterioration of the pore structure could not be unequivocally assigned to, for example, *Betula* or *Myrica*.

Rumex type

These trizonocolporate grains include species of *Rumex* and *Oxyria*. Because of their morphological similarity, no attempt was made to differentiate between the two genera.

Lycopodium

Six species of *Lycopodium* were identified, namely, *L. annotinum*, *L. clavatum*, *L. complanatum* type, *L. lucidulum*, *L. obscurum* and *L. selago*. Their differentiation was based primarily on type of sculptering, size of laesurae, grain shape and presence or absence of nodes (cf. McAndrews et al. 1973).

Filicales

Filicales includes all monolete, bean-shaped spores, that have a smooth surface and no perine, or are deteriorated. Spores in this category may belong to any of the following genera: *Athyrium*, *Asplenium*, *Cystopteris*, *Dryopteris*, *Onoclea*, *Polypodium*, *Polystichum*, *Thelypteris* and *Woodsia*.

Indeterminable

Pollen grains and spores in this category include those which cannot be confidently assigned to any taxon because of deterioration (i.e. corrosion, breakage, crumpling) or concealment (by extraneous material).

Unknown

Pollen and spores that are well preserved and possess distinctive features but for which no positive identification can be made are included in this category.

4.4 Approaches to Vegetation Reconstruction

The interpretation of the fossil pollen record begins with the determination of the taxa present, and their abundance at sequential levels (time periods) in the sediment profile. Past community reconstructions are then attempted using this information. With the exception of one approach utilizing direct evidence (e.g. preserved communities buried intact as a result of a sudden catastrophe), the methods of reconstructing past plant communities must be inferential by nature.

The most widely adopted approaches in Quaternary pollen studies are based on the use of modern analogues: present-day models extended backward in time.

One of these approaches involves the use of 'indicator species', defined as "morphologically distinctive pollen and spores of taxa of narrow ecological or sociological amplitudes today that are characteristic or diagnostic of particular present-day plant communities" (Birks [illegible] p.283). The method assumes that no significant change in the ecological requirements or sociological affinities of the taxon involved has occurred through time. Palaeoecological reconstructions are best made using a group of taxa of similar modern preferences rather than a single taxon. Whereas a change in the ecological amplitude of a species cannot be tested, it is less likely that the requirements and behaviour of a group of taxa occurring together would differ significantly through time (Birks & Birks 1980).

Another approach employed in the reconstruction of past vegetation types is based on the comparison of fossil pollen assemblages with modern pollen assemblages which characterize distinct vegetation types. When the comparison is successful, it is assumed that the vegetation which produced the fossil assemblage resembled that producing the modern assemblage. When no match between fossil and contemporary pollen assemblages can be made, it must be concluded that either the surface sample coverage is inadequate or there is no contemporary equivalent for the past vegetation (Lamb 1984). Previous research in Newfoundland (e.g. Mellars 1981; Macpherson 1982b) has demonstrated that the use of the comparative approach is not always feasible due primarily to the paucity of local studies of modern pollen-vegetation relationships and the poor representation of certain taxa in the pollen rain. This latter problem has been iterated by numerous researchers working in the province. For example, larch (*Larix laricina*) is common in low productive forests in the northern Avalon Peninsula area (Macpherson 1982b), yet its poorly preserved pollen rarely shows up in the pollen record. Similarly, *Juniperus communis* and members of the heath (Ericaceae) family, including *Ledum groenlandicum* and *Vaccinium*, are vastly underrepresented in modern pollen spectra (Morrison 1970; Mott 1974) hindering the differentiation of vegetation types. In addition, *Cladonia*, an important constituent of open lichen woodland and a main component of lichen heath vegetation types, which together cover large areas in Labrador and to a lesser extent in Newfoundland, is not represented in the pollen record. Such 'blind spots' are the greatest cause of difficulty encountered in distinguishing the boreal woodland and forest-tundra zones. As yet, the most reliable method of distinguishing the two zones is on the basis of total pollen influx.

The provenance of modern surface samples constitutes another limitation to the use of the comparative approach in this region. Whereas most fossil spectra are derived from lake sediments or peat bogs, surface samples are obtained from moss polsters in addition to bogs and lake deposits. Modern pollen spectra from moss surfaces are generally overrepresentative of local species. Lake sediment spectra, in contrast, comprise a mixture of pollen from all surrounding communities and tend to characterize the regional pollen rain (Webb et al. 1978).

Therefore, based on the experience of previous studies in the region, a variety of approaches to vegetation reconstruction has been employed in this study, namely the comparative approach using pollen percentages and pollen influx values, the use of indicator species, and reference to published interpretations of other fossil pollen assemblages which lack modern analogues.

CHAPTER 5 REGIONAL SETTING

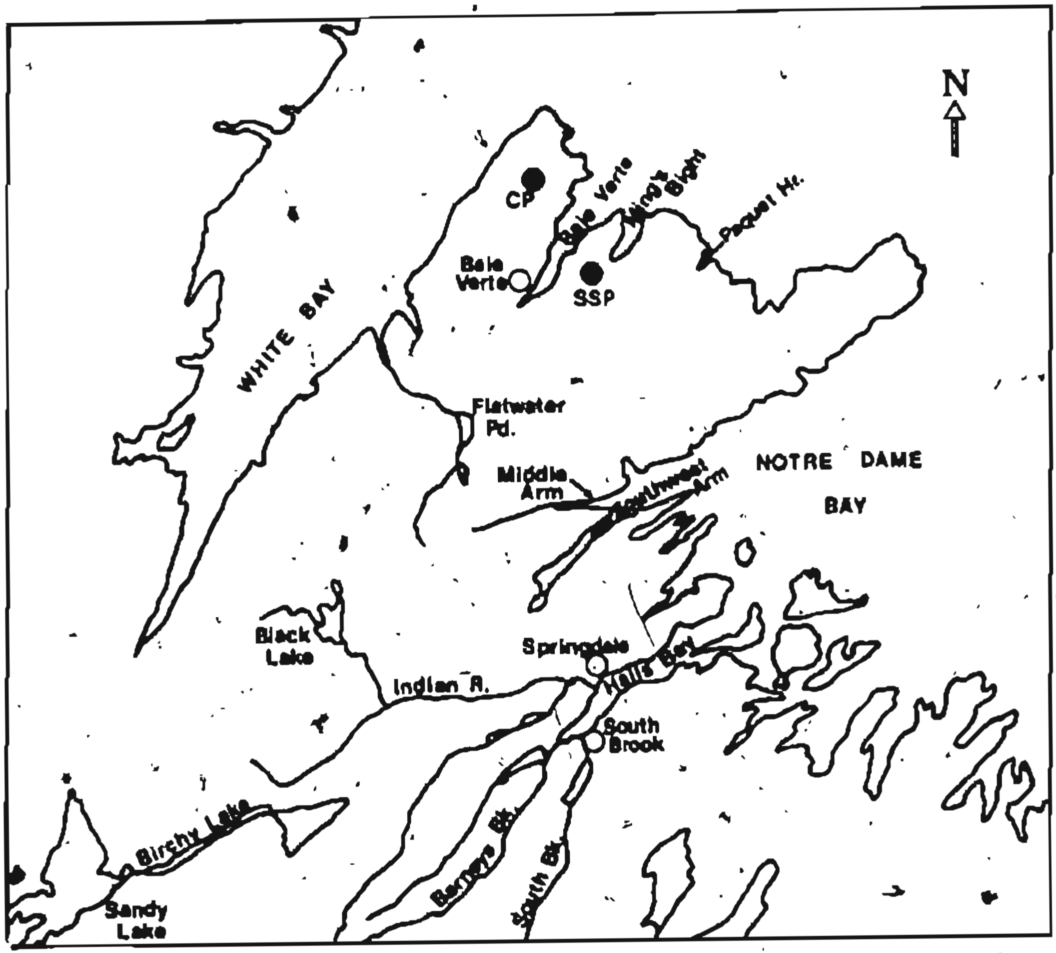
5.1 Location

The Baie Verte Peninsula (formerly the Burlington Peninsula), approximately 500km^2 in area, is located in northcentral Newfoundland and projects northeastward into the Atlantic Ocean (see Figures 5.1 & 5.2). It is bounded to the east by Green Bay and Notre Dame Bay, and to the west by White Bay. Hibbard (1983,1984) regards the peninsula's southern boundary to be the major fault-controlled valley connecting Southwest Arm (49°34'N, 56°10'W) of Green Bay with Birchy Lake (49°17'N, 56°55'W). Three smaller peninsulas are contained within the Baie Verte Peninsula, namely the Cape St. John Peninsula in the northeast, the Point Rousse Peninsula between Ming's Bight and Baie Verte, and the Fleur de Lys Peninsula between Baie Verte and White Bay.

5.2 Geology

Strong (1984) suggests that the geology of the Baie Verte Peninsula (see Figure 5.2) has received more detailed study than any other part of the province. This interest reflects both its economic and academic importance. The peninsula has great mineral potential, and the exploration and production of this resource has continued since 1864. Indeed, it appears that far earlier than this (ca. 2000 yrs BP), the Dorset people utilized the soapstone deposit at Fleur de Lys (Nagle 1981). In addition, the position of the peninsula at the boundary of the ancient North American continent and the Iapetus Ocean has provided a testing ground for the application of plate tectonic theories (Hibbard 1983).

A detailed report on the geology of the Baie Verte Peninsula has recently been published by Hibbard (1983). It is largely from this source that the following paragraphs are derived.



CP Compass Pond
SSP Small Scrape Pond

Figure 5.1: Location map, Baie Verte Peninsula and environs.

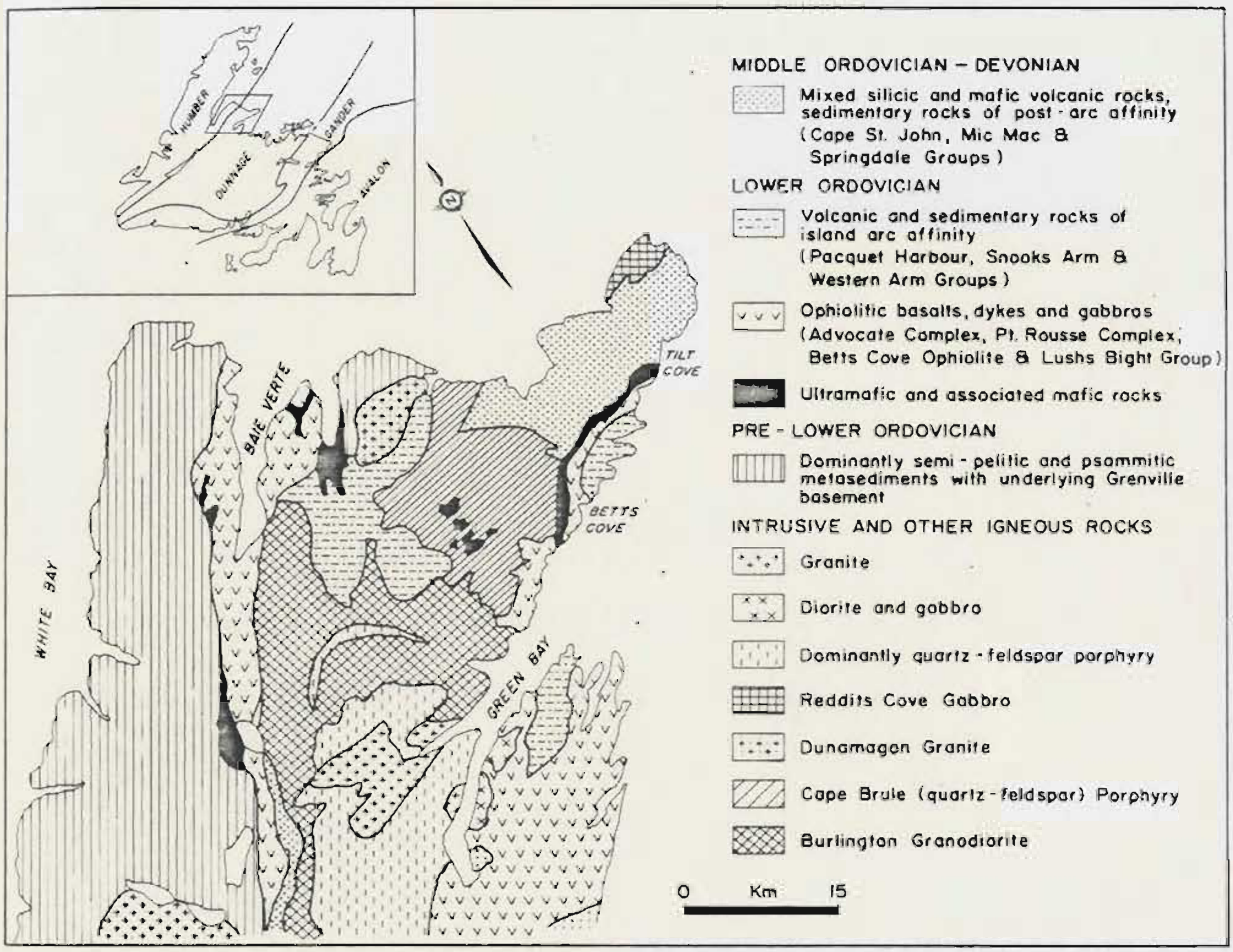


Figure 5.2: Geology of the Baie Verte Peninsula.

(Source: Strong 1984)

The Baie Verte Peninsula is situated at the northern terminus of the Appalachian geological and physiographic regions. It forms part of the Atlantic Uplands Division of the Canadian Appalachian geomorphic region; a gently southeasterly dipping upland of rolling terrain (King 1972). The hills in the west and southeast parts of the peninsula reach 300-350m.a.s.l. and the lowlands between, including the Cape St. John Peninsula, range from 175-200m.a.s.l.

The peninsula is bisected by a sharp structural zone termed the Baie Verte Line which trends north-northeast from Birchy Lake to Baie Verte, then east from Baie Verte to Paquet Harbour. This feature separates two tectonic zones: the Humber zone to the west and the Dunnage zone to the east. The eastern part of the Humber Zone, called the Fleur de Lys Belt, represents the eastern margin of the Late Hadrynian to Early Paleozoic North American continent. The belt is largely submerged beneath the Atlantic Ocean, north of the Baie Verte Peninsula. The southeastern portion of the peninsula forms the Baie Verte Belt and represents the western remnant of the Early Paleozoic Iapetus Ocean.

The Fleur de Lys Belt comprises three major lithic components, namely a local structural basement, a dominantly metaclastic cover sequence, and granitoid intrusions. The core of this belt, consisting of migmatites, banded gneisses, metaconglomerate, and psammitic and semipelitic schists, is Hadrynian and possibly older. The cover sequence, named the Fleur de Lys Supergroup, is composed of four groups which consist of psammitic, semipelitic and graphitic schist, marble, greenschist, and amphibolite. Regional correlation suggests the supergroup to be of Late Hadrynian to Early Ordovician age. Both the structural basement and cover sequence were post-tectonically intruded by granite.

The three major elements of the Baie Verte Belt comprise ophiolitic suites, volcanic cover sequences and intrusive rocks. The ophiolitic units are Early Ordovician or older and the cover sequences range in age from Early Ordovician to Devonian. Various intrusive rocks cut the ophiolites and volcanic complexes during the Ordovician and the Silurian and Devonian periods. At the southern end of the belt sedimentary rocks, probably of Carboniferous age, are exposed.

5.3 Glacial Geomorphology

The entire Baie Verte Peninsula was glaciated during the Pleistocene and most researchers consider the glacial deposits and many glacial markings to be related to Wisconsin glaciation. However, opinion is divided as to whether these features (see Figure 5.3) are of Late Wisconsin origin (e.g. Hibbard 1983) or older (e.g. Grant 1977b; Rogerson 1982).

Early workers on the peninsula (MacClintock & Twenhofel 1940) reported a sequence of till beneath and overlying basal marine clay at the head of Baie Verte which they claimed paralleled a similar sequence indicating ice retreat and readvance at Bay St. George, southwestern Newfoundland. Similarly, deltaic deposits with basal marine clays and overlying glacial outwash have been found near the heads of Southwest Arm and Middle Arm (Neale & Nash 1963). Only some of the observed fossiliferous material has been dated (see Table 5.1).

More information is available on the direction of glacial movement on the peninsula in the form of striae, grooves, roches moutonnees, crag and tail features and drumlinoid features (Hibbard 1983). In general, this evidence indicates that ice flowed northward from the interior of the island and radiated out from the centre of the peninsula (Hibbard 1983). The lowlands between Flat Water Pond

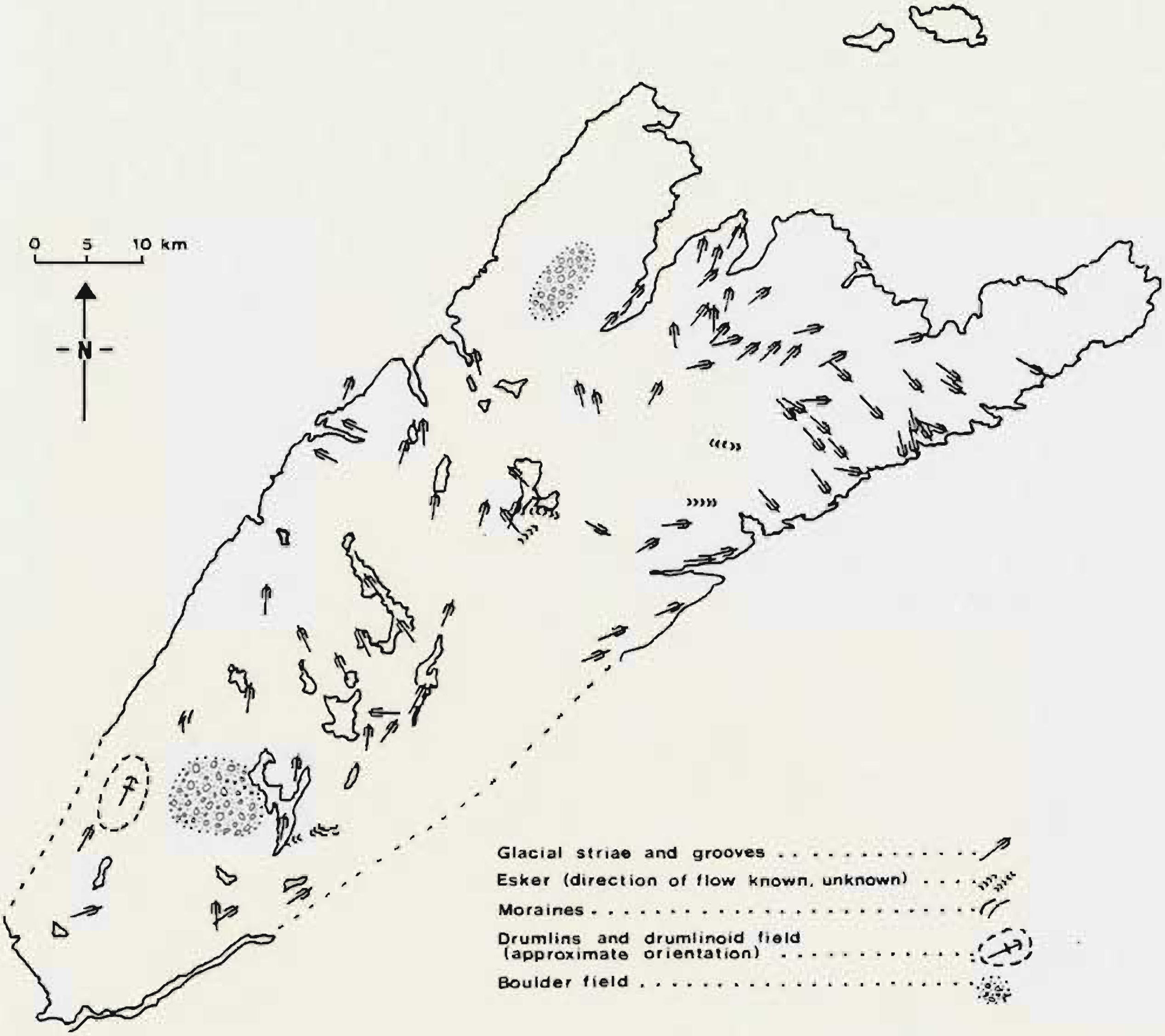


Figure 5.3: Glacial features of the Baie Verte Peninsula. (redrawn from Hibbard 1983).

and Birchy Lake served as an ice divide (Neale & Nash 1963); to the east ice flowed to Green Bay, and to the west ice flowed to White Bay. Lundqvist (1965), in describing sets of intersecting striae on the peninsula, suggested that the finer and younger striae resulted from local movements within a stagnating ice cap. Supporting this theory is the occurrence of hummocky ablation moraine in lowland areas between Sandy Lake and Flat Water Pond (Lundqvist 1965).

According to Hibbard (1983), the majority of boulders in till overlying large granitic bodies is derived from local sources. Erratics from the central ultramafic belt have been found west of Black Lake and east to Southwest Arm (Neale & Nash 1963), and attest to the concept of a central ice divide. In contrast, an erratic discovered on the west coast, near Shoal Point, has no local source and may have originated in the Long Range Mountains or in Labrador (Hibbard 1983).

Emergence features around the coast of the peninsula are indicative of postglacial uplift. The highest recorded raised shoreline is a gravel terrace at Seal Cove, White Bay, at 65m (Henderson in Neale & Nash 1963).

5.4 Soils

Orthic Humo-Ferric Podzols and Gleyed Humo-Ferric Podzols, developed in morainal materials, are the characteristic soils of the Baie Verte Peninsula. The former group are distinguished by a light grey surface horizon (Ae) of approximately 10cm depth overlying a reddish-brown horizon (Bf) of about 20cm depth in which Fe, Al, and organic matter accumulate (Guthrie 1981).

The Gleyed Humo-Ferric Podzols are similar, comprising a thick moss layer (Lf) and approximately 10cm of dull grey surface horizon (Ae) overlying a brown to dark brown horizon (Bfg) approximately 20cm thick. Both the Orthic and Gleyed Humo-Ferric Podzols produce fair to good forest growth (Guthrie 1981).

In addition, Ortstein Humo-Ferric Podzols, developed in morainal deposits, are found on the peninsula but are restricted to the southwestern region near Hampden. The properties of these soils are similar to the Humo-Ferric Podzol great group except that the B horizon is strongly cemented. Because of this hardening, the soils support either poor forest growth or ericaceous shrub vegetation. Southwest of Southwest Arm is a minor occurrence of Ortstein Humo-Ferric Podzols, developed in glacio-fluvial materials, which similarly support only poor forest growth (Guthrie 1981).

5.5 Climate

According to Banfield (1981), the Baie Verte Peninsula spans two climatic zones (see Figure 5.4). The climate of the northern half is oceanic and forms part of the East coast and hinterlands zone. The southern half forms part of the Central lowlands zone which has the most continental climate of the island. Tables 5.1 and 5.2 provide a summary of available climatic data for the region. Data for the Springdale station, to the east of the peninsula, is included to provide better coverage.

The climate of the northern part of the peninsula is influenced by the cold Labrador current. Southward-drifting pack ice is usually present along the northcentral coast until late in May and can be present until late June (Farmer 1981). Yet the frost-free period is slightly longer in the north, averaging 97 days at Baie Verte and 82 days at Springdale (Env.Cda. 1982a). The average date for the last spring frost is June 10th and that of the first fall frost is September 16th at Baie Verte, whereas June 14th and September 5th mark the beginning and end of the frost-free period at Springdale (Env.Cda. 1982a). In addition, night frosts

Figure 5.4: Climatic zones of Newfoundland.

(Source: Banfield 1983)
(key on following page)

Figure 5.4 cont'd

1. South and south-east coasts and immediate hinterlands.

2. Central uplands.

3. East coast and hinterlands. Less wet than 1 (900-1200mm per year). Occasional heavy precipitation with northeasterly or northerly airflow, especially in Gander area. Winters cold with 50-70% precipitation falling as snow; snow cover normally continuous at least throughout February (longer to the northwest). Cool, late springs with sea ice persisting until mid-May. Summers generally warm and fairly sunny.

4. Central lowlands. Greatest continentality. Annual precipitation 900-1100mm. Winters colder and drier than 1 and 2, with 65-75% precipitation falling as snow; occasional very severe frost (minima near -25 to -30°C), especially in Exploits and Humber Valleys and near Green Bay. Spring arrives relatively earlier in Humber Valley. Summers warm and moderately sunny, with maxima often 26-32°C during SW-SE airflow. Least windy and least foggy area.

5. West coast (Bonne Bay and south).

6. Northern Peninsula.

(see Banfield 1983 for full details on each climatic zone.)

Table 5.1 Climate Data for Baie Verte, Newfoundland
$49^0 59'$N $56^0 11'$W 110m

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Year
Daily Max.Temp	-3.6	-3.7	0.0	3.6	9.0	16.5	21.2	19.8	15.4	9.3	4.4	-0.9	7.6
Daily Min.Temp	-12.7	-14.4	-9.7	-4.1	0.4	5.4	10.0	9.5	5.5	0.9	-2.6	-9.0	-1.7
Daily Temp.	-8.2	-9.1	-4.8	-0.2	4.8	11.0	15.6	14.7	10.5	5.1	0.9	-5.0	2.9
Rainfall	30.4	14.0	29.8	24.9	69.2	74.4	69.1	95.7	94.4	85.4	80.6	32.9	700.8
Snowfall	75.6	72.7	60.8	38.6	9.8	0.1	0.0	0.0	0.0	5.0	30.0	77.1	369.7
Total Precip.	104.6	83.3	82.9	63.5	69.0	79.4	69.1	95.7	94.4	100.7	112.0	110.1	1064.7
Days with rain	2	2	3	4	10	11	10	12	14	13	8	3	92
Days with Snow	10	8	10	6	2	0	0	0	0	1	4	9	50
Days with Precip.	12	10	11	10	10	11	10	12	13	13	12	12	136
Degree Days Above 5^0C	0.1	0.2	0.5	3.4	41.9	175.6	332.4	300.6	167.9	46.0	10.2	1.3	1080.1

Units: Temp °C
Rainfall mm
Snowfall cm
Total Precipitation mm

Source: Canadian Climate Normals, 1951-1980.

Table 5.2 Climate Data for Springdale, Newfoundland
49°30'N 56°5'W 23m

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec	Year
Daily Max.Temp	-2.4	-2.6	1.2	5.2	11.4	18.0	22.5	21.3	16.6	10.5	5.4	-0.2	8.9
Daily Min.Temp	-12.5	-13.4	-8.7	-3.0	1.3	5.8	11.0	10.3	5.6	1.6	-2.1	-9.1	-1.1
Daily Temp	-7.5	-8.0	-3.8	1.1	6.3	11.9	16.8	15.8	11.1	6.1	1.7	-4.7	3.9
Rainfall	27.6	12.7	34.6	31.7	64.9	76.8	71.6	92.6	87.9	98.9	79.5	39.4	718.2
Snowfall	57.2	53.8	43.1	21.8	1.5	T	0.0	0.0	0.0	2.7	14.3	49.7	244.2
Total Precip.	87.9	67.3	77.7	54.4	66.4	76.8	71.6	92.6	87.9	101.7	93.8	89.0	967.1
Days with Rain	3	2	3	4	9	10	9	11	10	11	8	4	84
Days with Snow	7	6	6	4	*	0	0	0	0	*	2	6	31
Days with Precip.	9	7	9	8	9	10	9	11	10	11	10	10	113
Degree Days Above 5°C	0.2	0.2	1.2	6.1	66.6	210.3	366.5	335.7	185.4	65.3	16.8	2.3	1256.6

Units: Temp °C
Rainfall mm
Snowfall cm
Total Precipitation mm
T Trace amount, i.e. <0.1mm of liquid precipitation or <0.1cm of snowfall frozen precipitation.
* amounts <0.5 except zero

Source: Canadian Climate Normals, 1951-1980.

often occur in the early part of the vegetative season in the south (Damman 1983).

The mean temperature of the warmest month is 15.6°C at Baie Verte (Env.Cda. 1982a), but when northerly and easterly winds prevail, the northern area experiences sudden cold spells in the summer (Damman 1983). The mean temperature of the warmest month at Springdale is 16.8°C (Env.Cda. 1982a). The southern part of the peninsula is also one of the least windy and least foggy areas of the province (Banfield 1981). The number of degree-days above 5°C varies little between the two areas, being slightly over 1200 in the south and between 1100 and 1200 in the north (Banfield 1981; Env.Cda. 1982a).

The approximate length of the vegetative season (based on a threshold of 5.5°C) of the Baie Verte Peninsula is 150 days. During this period the northern half experiences slightly lower mean temperatures (13-14°C) than does the southern half (15.5-16°C)[1] (Damman 1976). The peninsula receives somewhat less precipitation than the rest of the island ($<$1100mm annually) (Env.Cda. 1982a) and this, coupled with warm summers and high evapotranspiration losses (Damman 1983), makes it one of the driest parts of the island.

Winters are cold throughout the region; the mean temperature of the coldest month reaches -9.1°C at Baie Verte and -8.0°C at Springdale (Env.Cda 1982a). Undoubtedly the temperatures at these stations are tempered by their position at the heads of inlets, and higher inland areas probably experience far lower temperatures. The amount of winter precipitation falling as snow is similar for both areas; 50-70% in the north and 65-75% in the south (Banfield 1981).

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[1] Temperatures are grassland values based on sucrose inversion method

However, snow cover persists for a longer period in the north, and in the south, very severe frost is occasionally experienced (Banfield 1981).

5.6 Vegetation

The Baie Verte Peninsula lies in Forest Region 3 (Govt. of Nfld 1974) which encompasses northcentral Newfoundland. A breakdown of the land class areas for this region and percentages of gross merchantable volume are presented in Table 5.3. The table indicates *Picea mariana*, *Abies balsamea* and *Betula* (mainly *B. papyrifera*) as being the dominant forest tree species.

More recently, Damman (1983) has divided the island into ecoregions and the Baie Verte Peninsula bridges two such regions, namely Central-Newfoundland and the North Shore (see Figure 5.5). In the former ecoregion, *Abies balsamea* forests predominate in areas undisturbed by fire. Kalmia-black spruce and Pleurozium-balsam fir forests are also common (Damman 1964). In the northern part of this ecoregion *Picea mariana* replaces *Abies balsamea* as the dominant tree. This is a result of fires which, between 1958 and 1974, burned over 46,000 ha of forest land (Damman 1983). According to records extending back to the early 1600s (Wilton & Evans 1974), the Baie Verte Peninsula has experienced numerous forest fires, although the southern and central parts of the peninsula are most severely prone. Fire frequency in the northern part of this ecoregion has also, but to a lesser degree, promoted the growth of *Betula papyrifera* and *Populus tremuloides* stands. In addition, frequent forest fires have often been responsible for the development of *Kalmia angustifolia*-dominated dwarf shrub heath communities on poor soils. Raised bogs occur throughout the Central Newfoundland ecoregion and low *Picea mariana* and *Larix laricina* often grow on the strings.

Table 5.3 Land Class Areas and Gross Merchantable Volume for Northcentral Newfoundland

	(000's of acres)	(Percentage)
Productive forest	1943.7	63,3
Hardwood scrub	32.9	1.1
Softwood scrub	411.9	13.4
Bog	312.7	10.2
Rock barren	102.9	3.4
Cleared land	33.4	1.1
Agricultural land	1.8	0.1
Water	230[illegible]	7.5

	(000's of cunits)	(Percentage)
Black spruce	10,536.5	52,0
White spruce	358,6	1.8
Balsam fir	5,560.7	27.4
White pine	115.4	0.6
Larch	412.4	2.0
Birch	2,348.8	11.6
Other hardwood	946.7	4,7

Source: Government of Newfoundland 1974

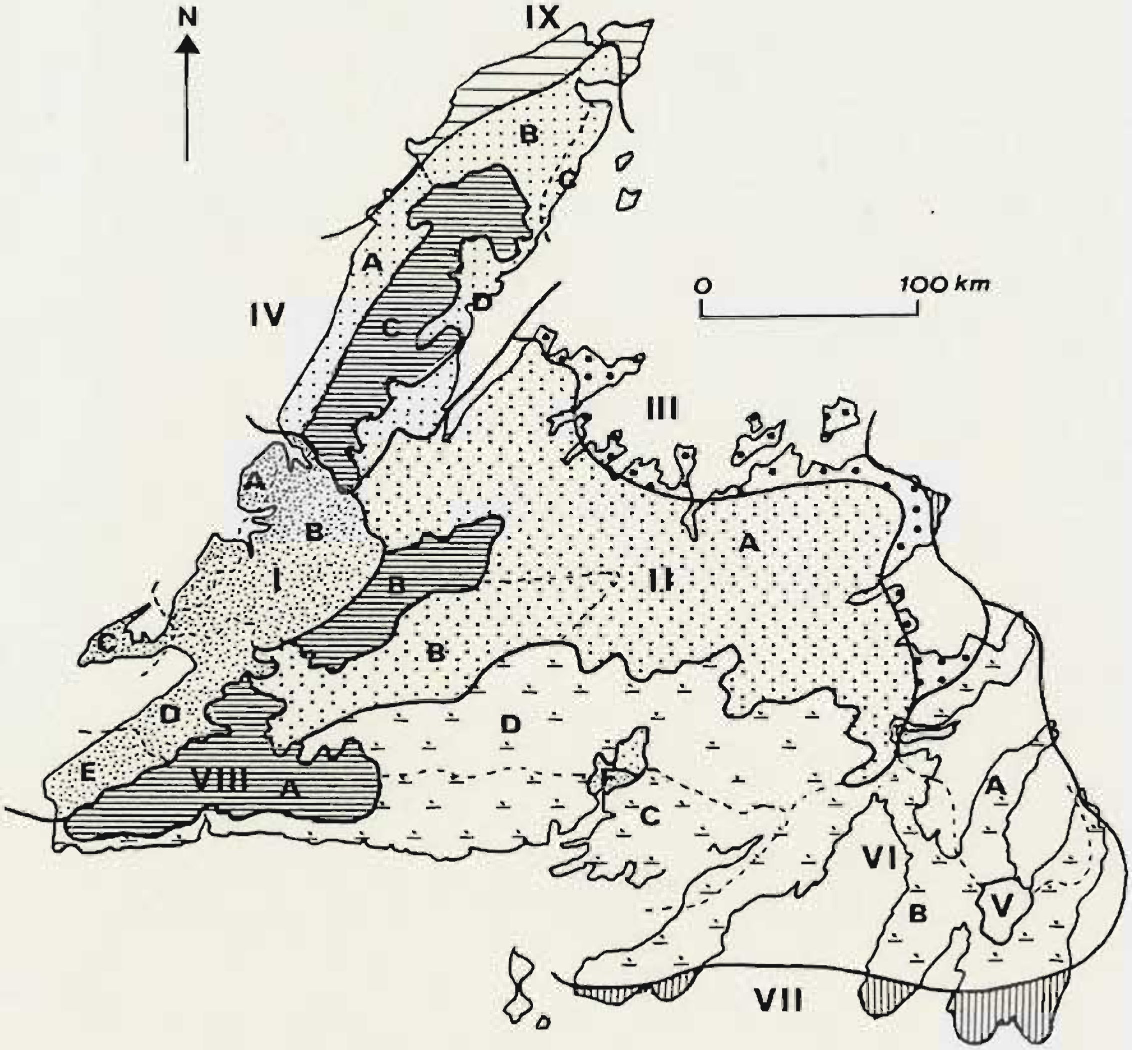

Figure 5.5: Ecoregions and subdivisions of the Island of Newfoundland. (redrawn from Damman 1983).

I=Western Newfoundland, II=Central Newfoundland, III=North Shore, IV=Northern Peninsula Forest, V=Avalon Forest, VI=Maritime Barrens, VII=Eastern Hyper-oceanic Barrens, VIII=Long Range Barrens, IX=Strait of Belle Isle. (see Damman 1983 for description of subregions)

To a large extent, the North Shore Ecoregion is similar to the Central Newfoundland Ecoregion. However, *Picea glauca* is more abundant, stands of *Populus tremuloides* are absent and *Alnus crispa* occurs on both dry and wet sites in the northern ecoregion. In addition, *Alnus crispa* predominates in the alder swamps, as *A. rugosa* is rare in the northern part of the Baie Verte Peninsula (Damman 1983).

Most arctic-alpine plants on the Baie Verte Peninsula are restricted to the hills. However, a few species, namely *Cornus suecica* (Northern Dwarf Cornel), *Empetrum atropurpureum* (Purple crowberry) and *Vaccinium uliginosum* (Bog bilberry) occur in the northern coastal areas (Damman 1983).

CHAPTER 6 SITE CHARACTERISTICS

6.1 Compass Pond

[i] Location

Compass Pond (50°02'03"N, 56°11'47"W, grid ref.5 580 55 425) is situated 4km from the eastern coast of the Fleur de Lys Peninsula and 12.2km north of the town of Baie Vert[illegible] Route 410 (see Figure 6.1). It is well above the estimated marine limit of 75 to 100m (Grant 1980), lying at an elevation of approximately 236m.

[ii] Topography

The lake occurs in an area of rolling terrain and summits of 320m and 260m lie to the southwest and southeast of the lake. Its catchment is underlain by ridged bedrock of schists with minor greenschists and marble of the Eocambrian to Lower Ordovician Rattling Brook Group, of the Fleur de Lys belt (Hibbard 1983), and is veneered with moraine (Guthrie 1981). Although the site lies just north of the Exploratory Soil Survey region (Guthrie 1981), it appears that Shallow Gleyed Humo-Ferric or Orthic Humo-Ferric Podzols are the dominant soil types.

[iii] Lake Characteristics and Drainage

This small headwater pond, with a surface area of ca. 3.5ha, measures approximately 150 x 350m with its longitudinal axis trending north-northeast (see Figure 6.1), reflecting the structure of the underlying bedrock and the path of glacial erosion (Hibbard 1983). One small outlet drains the lake to the northeast and the only inflow is provided by a small ditch draining a culvert at the southeast end. The lake's catchment is small, ca. 43ha. Coring was performed in

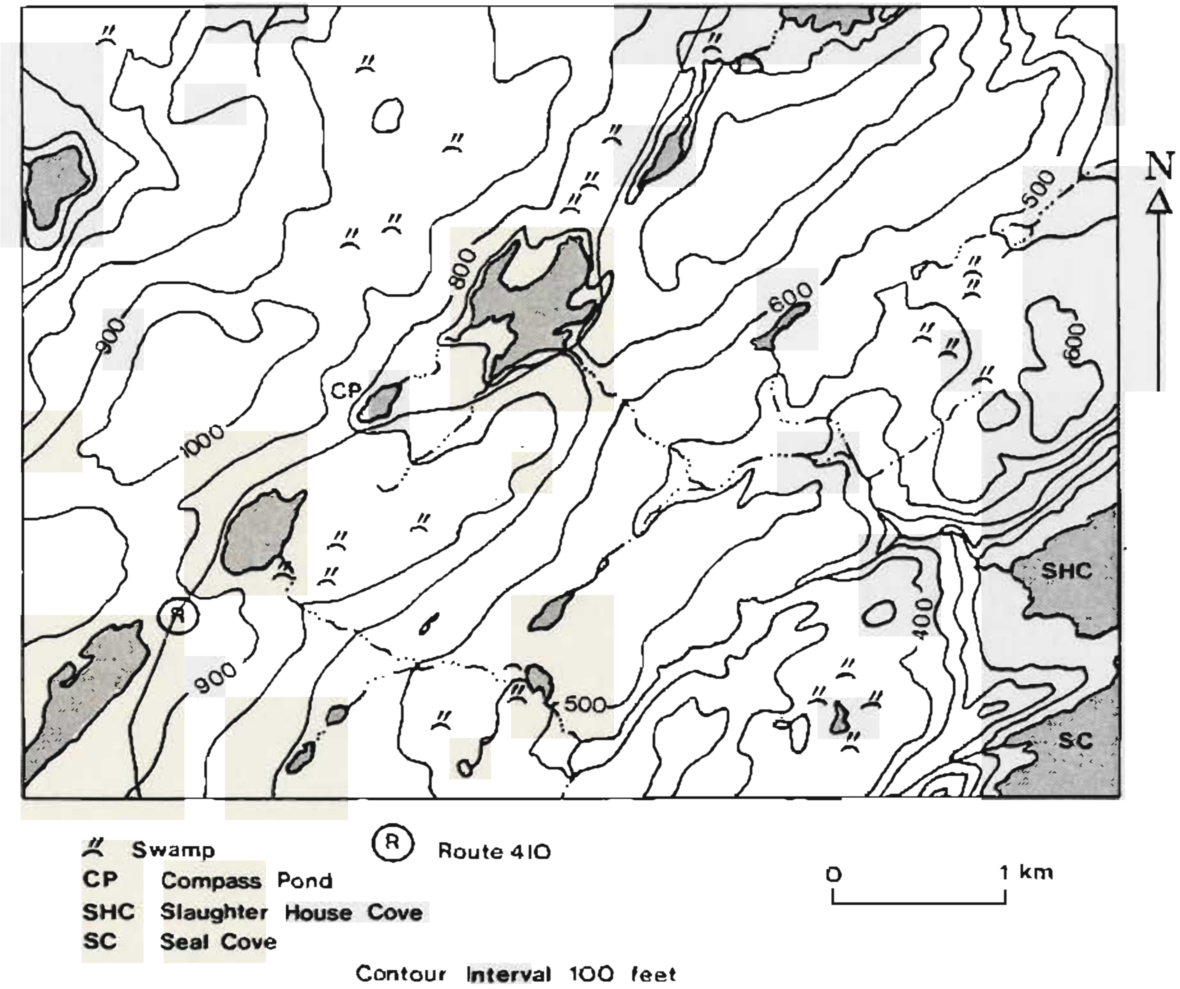

Figure 6.1: Topographic map of Compass Pond area.

the centre of the lake where water depths vary from 2-3m. A maximum depth of over 6m was recorded at the northern end while an extensive shallow area, less than 2m, occurs at the southern end.

[iv] Vegetation

Taxonomic nomenclature follows Gleason and Cronquist (1963). Abundant *Nuphar variegatum* and a small clump of *Lobelia Dortmanna* are the only aquatic plants supported by the lake. *Myrica gale* is abundant at the water's edge. Other shrubs immediately surrounding the lake include *Kalmia angustifolium*, *Ledum groenlandicum*, *Salix discolor*, *S. argyrocarpa*, *Chamaedaphne calyculata* and *Alnus crispa*.

At the southeast side of the lake, where the ditch enters, Cyperaceae (including *Scirpus rubrotinctus*) and Gramineae (including *Calamagrostis Pickeringii*) are abundant. *Sanguisorba canadensis* is also common, whereas *Aster puniceus* and *Habenaria dilatata* are infrequent. Sunny and damp areas adjacent to the lake favour the growth of *Pteridium aquilinum*, *Osmunda claytonia*, *Equisetum sylvaticum*, *Juncus effusus*, *Carex brunnescens*, *Vicia cracca* and several shrubs including *Salix* spp., *Cornus stolonifera* and *Rubus idaeus*. A small bog southeast of the lake supports Cyperaceae, Gramineae, *Sanguisorba canadensis*, *Alnus* and low *Picea mariana*.

In sunny and disturbed areas (along the roadside and on top of roadcuts), 'weedy' species are typically abundant, including *Taraxacum officinale*, *Achillea Millefolium*, *Trifolium*, *Anaphalis margaritacea* and *Epilobium angustifolium*. *Betula papyrifera* (and possibly *B. cordifolia*), and small trees of *Populus tremuloides*, *Sorbus*, *Prunus pensylvanica* and *Acer rubrum* occur at the forest edge together with *Salix* shrubs, *Alnus crispa*, *Cornus stolonifera* and the occasional *Ribes lacustre*.

Most of the area is covered with a fairly dense forest, similar to Damman's (1964) *Kalmio-Piceetum* forest type, comprising *Picea* and *Abies balsamea* with lesser amounts of *Betula papyrifera*. *Kalmia angustifolium* predominates in the shrub layer. *Rhododendron canadense*, *Ledum groenlandicum*, *Linnaea borealis*, *Gaultheria hispidula*, *Vaccinium angustifolium*, *V. ovalifolium*, *V. Vitis-Idaea* and *Sorbus* are also present. In addition, *Taxus canadensis*, *Acer spicatum*, *Viburnum edule* and *Aralia nudicaulis* occur in the understory in cool and shady lowland sites.

Mosses and *Cornus canadensis* profusely carpet the forest floor. The herb layer also contains *Clintonia borealis*, *Trientalis borealis* and, infrequently, *Maianthemum canadense*, *Galium triflorum* and *Solidago macrophylla*, the two latter species growing in cool damp sites. Several fern species, including *Athyrium Filix-femina*, *Gymnocarpium Dryopteris* and *Pteridium aquilinum*, are also quite common.

Several shrubs and herbs dominate the small cutover areas, namely *Vaccinium angustifolium*, *V. ovalifolium*, *V. Vitis-Idaea*, *Alnus crispa*, *Cornus canadensis*, *Clintonia borealis* and *Epilobium angustifolium*, while seedlings of *Abies balsamea* and *Betula papyrifera* abound.

Although Damman (1983) lists *Pinus strobus* and *Alnus rugosa* as growing in the North Shore Ecoregion, neither species was found near the site.

6.2 Small Scrape Pond

[i] Location

This lake (49°56'59"N, 56°05'17"W, grid ref.5 657 55 325) lies 4.5km southwest of Ming's Bight on Route 418 and is approximately 15km southeast of Compass Pond (see Figure 6.2). At an elevation of approximately 122m, it too lies well above Grant's (1980) estimated marine limit.

[ii] Topography

The site is in an area of rock knobs which reflects the complex bedrock structure of the Dunnage Tectonic Zone. The area to the southwest is underlain by mafic volcanic and volcaniclastic rocks and diabase dikes of the Middle Ordovician Paquet Harbour Group. Immediately northeast of the lake the hills, rising to about 170m, consist of serpentized ultramafic rock of the Point Rousse Complex of Late Cambrian to Early Ordovician age. The area to the north and east is underlain mainly by gabbro and metagabbro of the Point Rousse Complex (Hibbard 1983). Morainal veneer with bedrock exposure is characteristic of the area, and well-drained but shallow Orthic Humo-Ferric Podzols are the main soil types (Guthrie 1981).

[iii] Lake Characteristics and Drainage

Small Scrape Pond is an elongated water body, approximately 425 x 50m with a surface area of ca. 3.5ha. Its northwest trend is structurally controlled by a major fault, the Scrape Thrust Fault. It is a headwater pond draining northwest into Scrape Pond and has two small inlets: one at the southwest side, the other at the northwest side. The size of its catchment, at ca. 47ha, is similar to that of the Compass Pond. Traverses of the lake determined the deepest parts

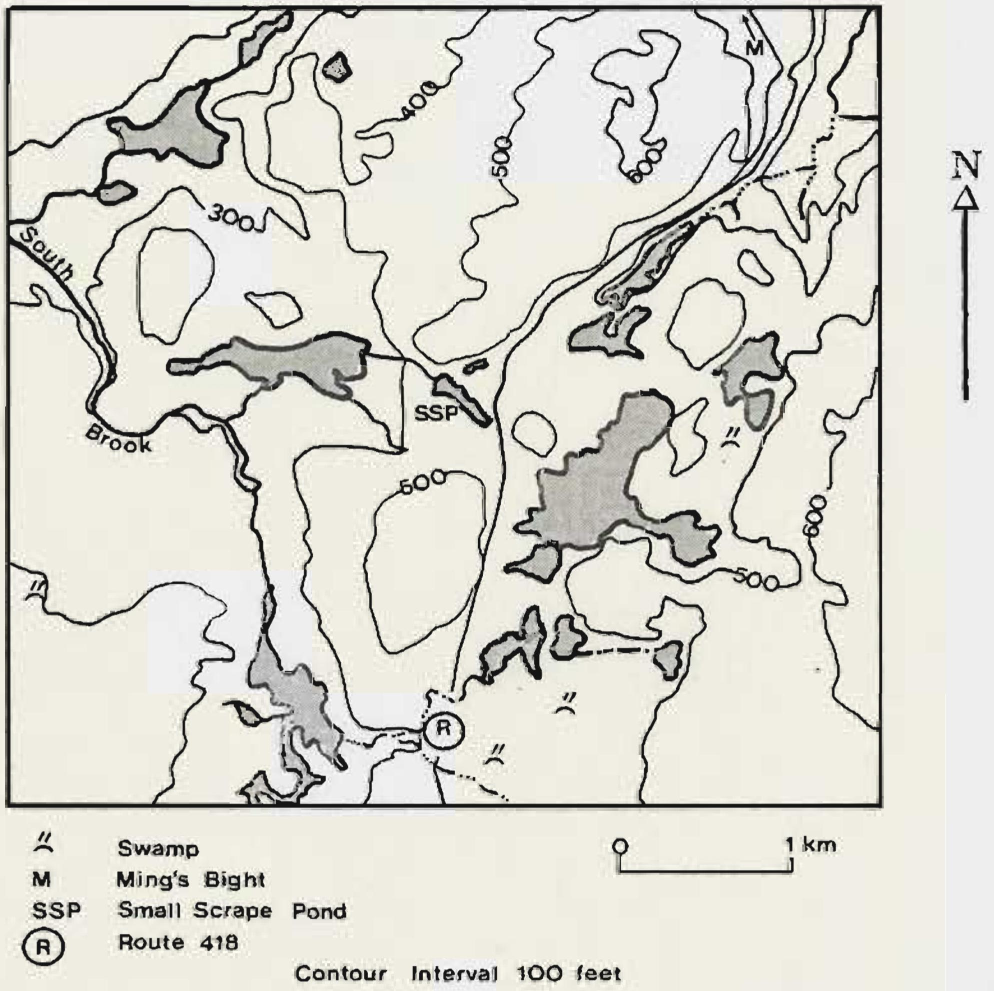

Figure 6.2: Topographic map of Small Scrape Pond area.

to be at the western end (over 6m) and just east of the island (about 7m) and the shallowest parts to be at the eastern end (2-3m) and just west of the island (3-4m). The two coring locations were in the western half of the lake in a water depth of 4.9m.

[iv] Vegetation

The only aquatic species growing in the shallows of Small Scrape Pond is *Nuphar variegatum*. Cyperaceae and Gramineae are abundant at the lake's edge as are *Alnus*, *Myrica gale*, *Salix*, *Sorbus*, *Chamaedaphne calyculata*, *Sanguisorba canadensis*, and clumps of *Potentilla fruticosa*.

The area to the south and west of the lake has been logged and the secondary growth consists mainly of *Betula papyrifera*, *Abies balsamea* and *Prunus pensylvanica*. To the northeast, the forest canopy is similar to that surrounding Compass Pond. The area supports a dense shrub layer dominated by *Ledum groenlandicum* and *Kalmia angustifolia*. *Vaccinium angustifolium*, *Gaultheria hispidula*, *Cornus stolonifera*, *Amelanchier bartramiana*, *V. Vitis-Idaea*, *V. ovalifolium* and *Rhododendron canadense* are also present. This nearly impenetrable shrub layer has resulted in an impoverished herb layer of which *Cornus canadensis* and *Clintonia borealis* are the most striking components. At sites where the shrub layer is less dense and near the edge of the woods, *Aralia nudicaulis*, *Actaea rubra*, *Solidago macrophylla*, *Aster umbellatus*, *Streptopus amplexifolius*, *Pteridium aquilinum* and *Thelypteris palustris* occur. *Taxus canadensis*, *Viburnum edule*, *Ribes glandulosum* and *Salix (mainly S. planifolia* are also encountered at the forest edge.

A few small bogs occur in the catchment. The following taxa are typical at these sites: *Betula michauxii*, *Picea mariana*, *Larix laricina*, *Alnus*, *Juniperus*, *Myrica gale*, *Andromeda glaucophylla*, *Potentilla fruticosa*, *Utricularia cornuta*, *Thalictrum*, *Triglochin maritima* and Cyperaceae including *Scirpus cyperinus* and *Carex michauxiana*.

The character of the vegetation changes gradually from the southern to northern parts of the Baie Verte Peninsula. Near the coast, the area is covered by poorly developed forests or barren lands. The two study sites, however, are located about 4km from the coast so that the effect of wind exposure is not a great factor in determining vegetational development. Compass Pond and Small Scrape Pond are situated near the boundary of two ecoregions (*sensu* Damman 1983). In terms of both composition and structure, the vegetation of the two sites is similar and may be regarded as typical of the Baie Verte Peninsula and, as such, the pollen collected in these lake basins in expected to be representative of the regional pollen rain.

CHAPTER 7 ANALYSIS OF LAKE SEDIMENTS

7.1 Compass Pond - Results

[i] Stratigraphy and Loss-on-ignition

The stratigraphy of the two Compass Pond cores is described in Tables 7.1 and 7.2. Both cores show the expected postglacial sequence of mineral sediment grading into organic sediment which is confirmed by the percentage loss-on-ignition (organic content) curve (see Figure 7.1).

The basal 71cm (565-636cm) and 73cm (547-620cm) of CPI and CPII, respectively, comprise a lower stiff grey clay layer grading into a thin layer of dark grey thixotropic silty clay. This unit is practically inorganic: the mean organic content is less than 5% in the light clay layer and increases to nearly 10% in the dark clay layer. Presumably, the layer was derived from the catchment slopes which supported a sparse vegetation cover. The presence of a pebble at 570cm in the CPI core is suggestive of ice-rafting.

The overlying clay-gyttja is 105cm (460-565cm) and 107cm (440-547cm) thick in the CPI and CPII cores, respectively. The presence of plant fibres within the sediment and the increase in loss-on-ignition, from 10% to 42% ($\bar{x}$ = 24.6%), signal the beginning of organic sedimentation. The upper sediments (0-440cm, CPI; 0-460cm, CPII) consisting of dark brown gyttja, have the highest organic content, ranging from 46.7% to 57% ($\bar{x}$ = 51.2%).

In general, the percentage increase in organic material closely parallels the rise in pollen concentration and is probably related to an increase in terrestrial biomass productivity. The presence of macrofossils, including leaves and twigs, at the top of the clay-gyttja supports this interpretation. However, the small bulge

Table 7.1 Stratigraphy of Compass Pond II

Depth (cm)	Sediment type	Munsell Colour	Comments
0-160	dark brown gyttja	10YR2/2	very wet & sloppy, often with plant fibres
160-395	dark brown gyttja	10YR2/2	wet, somewhat firmer, some plant fibres particularly at 310cm
395-440	dark brown gyttja	10YR2/2	firmer, deciduous leaves & twigs at 438cm
440-485	greyish brown clay-gyttja	2.5Y3/2	few plant fibres
485-547	dark olive grey clay-gyttja	5Y3/2	firm [illegible] micaceous
547-555	dark grey silty clay		thixotropic
555-557	medium grey silty clay		thixotropic, gritty, micaceous
557-601	medium light grey clay		thixotropic, firmer at 580cm, micaceous
601-605	light grey clay		soft, gritty
605-620	light grey clay		stiff, gritty

Table 7.2 Stratigraphy of Compass Pond I

Depth (cm)	Sediment type	Comments
0-460	dark brown gyttja	very wet & sloppy, often with plant fibres
460-565	clay-gyttja	leaves & twigs at 460-470cm, Myrica gale leaves & twigs at 460cm, 4 Picea needles & M.gale bract at 465-470cm.
565-571	dark grey clay	large stone at 570cm
571-636	light grey clay	sand at base

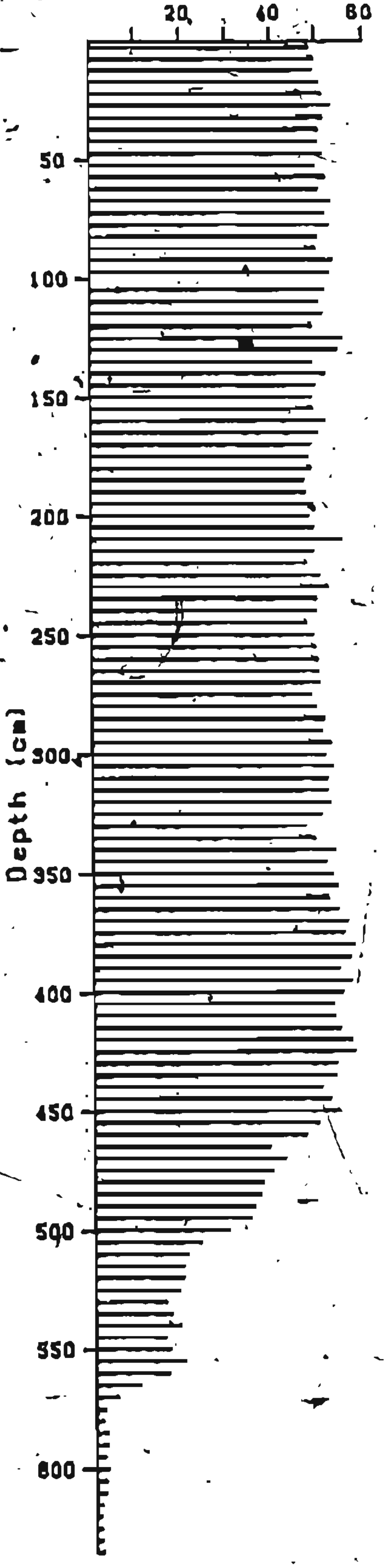

Figure 7.1: Compass Pond II, Percentage loss-on-ignition.

in the loss-on-ignition curve, near the base of the clay-gyttja, may represent an increase in lake productivity as attested to by an abundance of *Pediastrum* (observed in CPII), particularly in the lower half of this sediment unit.

[ii] Radiocarbon dates and sedimentation accumulation

Radiocarbon dates, based on a ^{14}C half-life of 5568±30 years and corrected for isotopic fractionation, δ^{13}C, are quoted in years before present (BP:1950AD). Table 7.3 summarizes the depths and radiocarbon ages from the Compass Pond II core from which a time-depth curve (see Figure 7.2) was established. The sediment deposition rates were calculated, using the POLSTA program, by joining the points for the mean age and the mean depth of each dated sample. Although this method produces mean sedimentation rates between dated levels, POLSTA 'smooths out' abrupt changes in the calculated sedimentation rates on either side of the dated levels.

For the Compass Pond II core, the rates of sediment accumulation show an overall increase from a mean of 0.312mm yr^{-1} in the clay-gyttja section to a mean of 0.581mm yr^{-1} in the gyttja section (see Table 7.3). The lowest segment of the core is undatable, due to its highly inorganic composition. However, the POLSTA program automatically calculated an extrapolated sedimentation rate for the lower segment (i.e. 0.257mm yr^{-1}). Fairly stable rates of sedimentation were attained during the early Holocene and the middle Holocene except for a decrease between ca. 8400 and 6300 BP. A second decrease in sediment accumulation occurred during the late Holocene (after ca. 2000 BP).

Table 7.3 Radiocarbon Dates for Compass Pond II

Depth below sediment surface (cm)	Lab. No.	Corrected & (Uncorrected) Age	$\delta^{13}C$	Dry Sample Wt(g).	Sedimentation rate(mm yr^{-1})
					0.471
95-100	GSC-3910	2050±90 (2060±90)	-25.5	3.8	
					1.000
195-200	GSC-3906	3050±140 (3080±140)	-26.82	2.7	
					0.610
295-300	GSC-3903	n/a (4690±160)	n/a	2.2	
					0.629
395-400	GSC-3902	6280±120 (6280±120)	-25.1	4.0	
					0.197
435-440	GSC-3992	8310±140 (8310±140)	-25.2	4.8	
					0.366
495-500	GSC-3898	9950±150 (9890±150)	-21.0	12.1	
					0.257
540-545	GSC-3891	11700±180 (11600±180)	-17.4	19.6	

Pretreatment of samples: NaOH omitted, HCl (no reaction), & distilled H_2O. Dates for GSC-3910, 3906 and 3903 were based on two 1-day counts in the 2L counter. Dates for GSC-3902, 3992, 3898 and 3891 were based on one 3-day count in the 2L counter. Sample GSC-3992 was mixed with dead gas for counting.

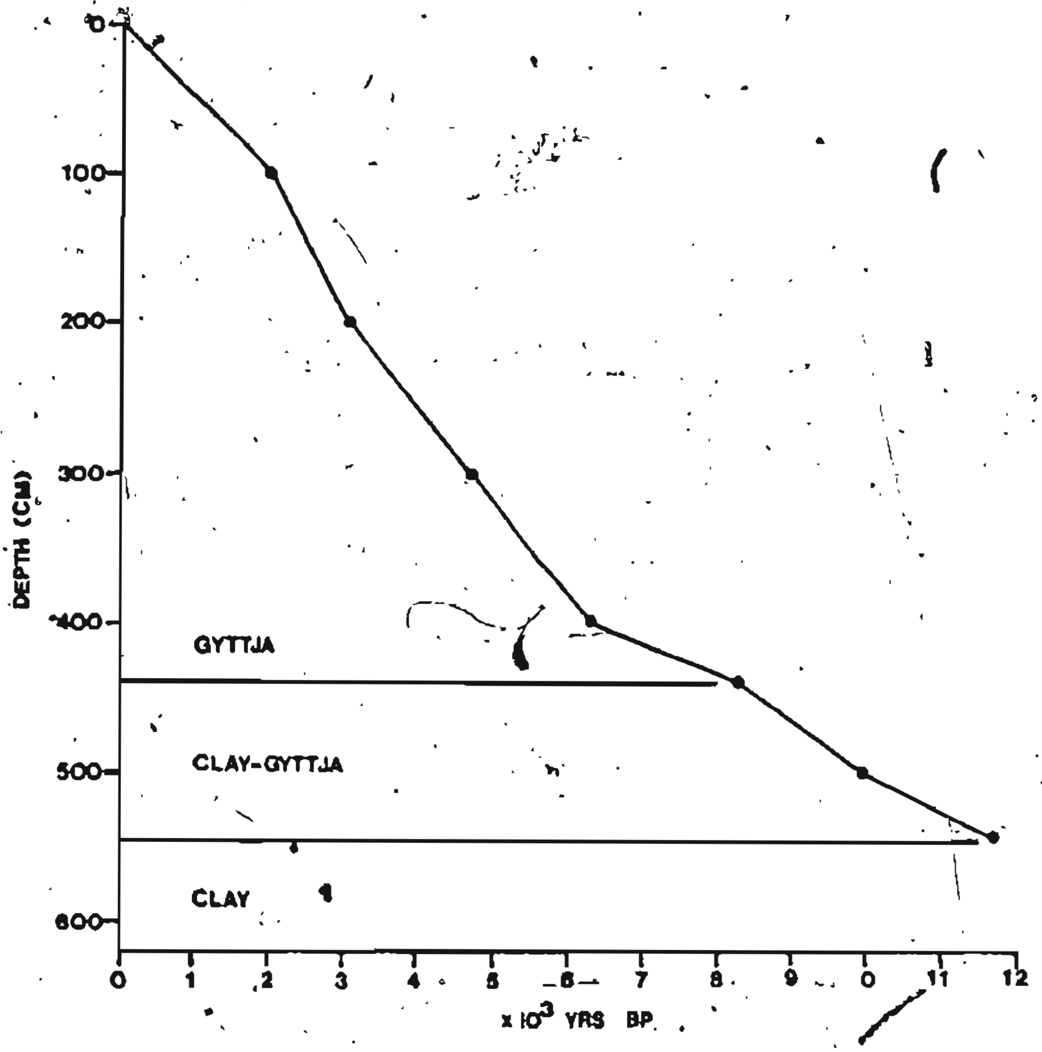

Figure 7.2: Compass Pond II, Time-depth curve.

[iii] Pollen zonation

The pollen diagrams for CPII (see Figures A.1, A.2, A.3, 7.3a,b) are divided into local pollen assemblage zones as outlined in Chapter 4 and are numbered and defined from the base upwards. The letters prefixing the zones refer to the site. Total pollen concentration and total pollen influx are based on those taxa comprising the pollen sum and, therefore, exclude spores, aquatics and *Pediastrum*.

A summary of the pollen and spore types and macrofossils encountered during the analysis is given in Table 7.4. An asterisk indicates grains and spores identified with the use of taxonomic keys only and of which, therefore, the determination is not wholly reliable.

Pollen assemblage zone CP-1: Gramineae-*Artemisia*-herbs (below 545cm; ended ca. 11,800 BP).

The pollen of herbs and extra-regional *Pinus* dominate this zone, the former contributing from 19% to 72%. In particular, the zone records successive maximum proportions of *Sphagnum* (up to 26%), Gramineae (up to 32%) and *Artemisia* (up to 24%). Other taxa, including Cyperaceae, *Ambrosia*, *Lycopodium* and Filicales occur consistently. Ericales also attain highest percentages (5%) in this zone and *Juniperus* peaks (at 17%) near the upper zone boundary extrapolated at ca. 12,000 BP. With the exception of one *Eriocaulon* grain at the basal level, pollen of aquatic taxa is absent from this zone. *Pediastrum* is present only at the base and near the top of CP-1.

There are smaller peaks of some herbs, namely Compositae (Tubuliflorae), Caryophyllaceae, Chenopodiaceae, Cruciferae and Leguminosae (the two latter

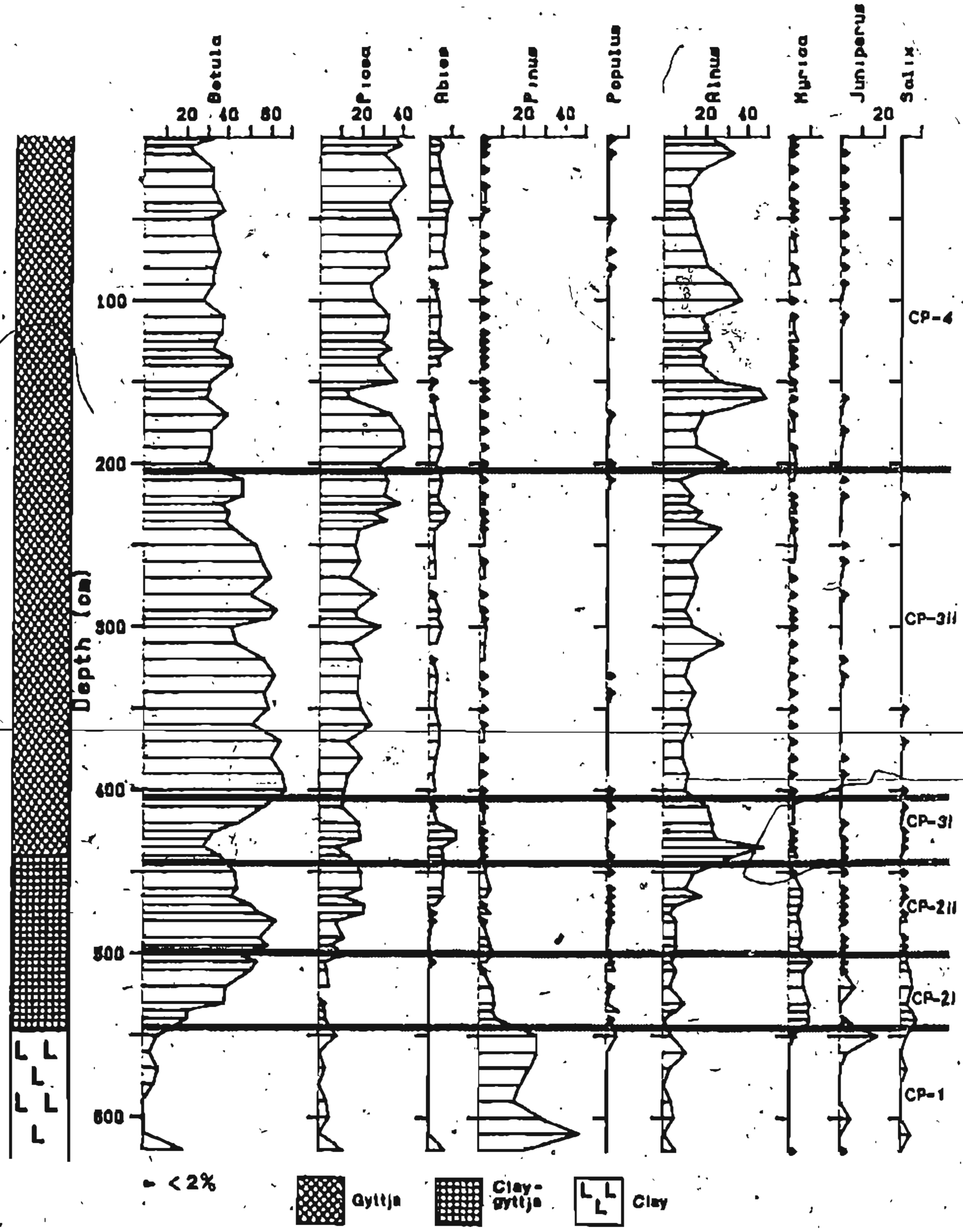

Figure 7.3a: Compass Pond II, Summary pollen percentage diagram, trees and shrubs.

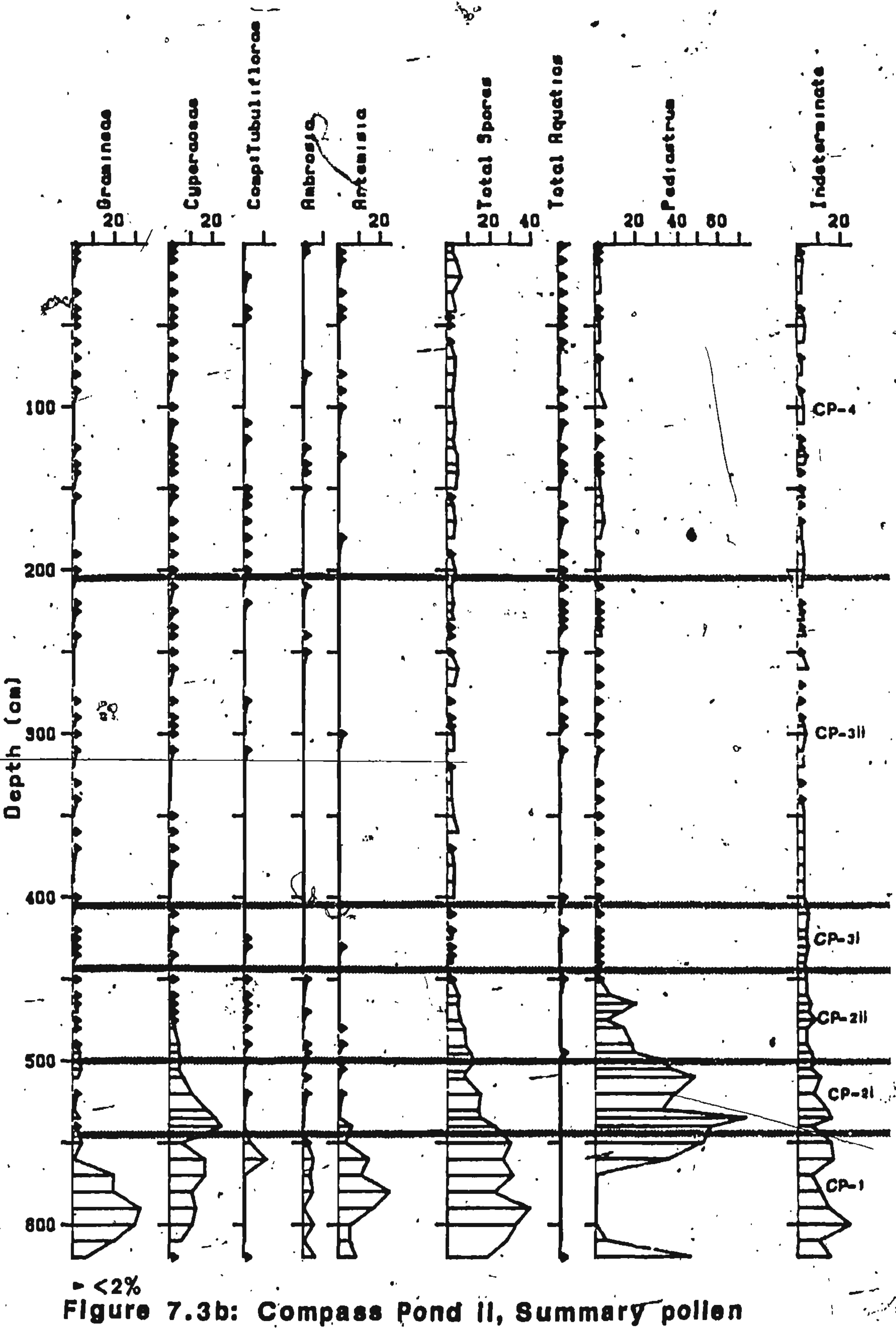

Figure 7.3b: Compass Pond II, Summary pollen percentage diagram, NAP.

Table 7.4 Fossil Pollen Grain and Spore Types Identified

Trees	CP site	SSP site
Abies balsamifera	+	+
Acer	+	+
*A.spicatum**	+	0
Betula	+	+
Carya	+	0
Fagus	+	0
Fraxinus	+	0
Juglans	0	+
Larix laricina	+	0
Picea @	+	+
P.mariana	+	+
P.glauca	+	+
Pinus	+	+
P.resinosa type	+	+
P.strobus	+	+
Populus	+	+
Quercus	+	+
Tilia	+	+
Tsuga	+	+
Ulmus	+	+
Shrubs		
Alnus	+	+
Coryloid type	+	+
Corylus	+	0
Cornus stolonifera	+	0
Ericales	+	+
Juniperus	+	+
Myrica @	+	+
Nemopanthus	+	0
Salix	+	+
Shepherdia canadensis	+	+
Taxus	+	+
Herbs		
Caryophyllaceae	+	0
Chenopodiaceae	+	+
Compositae	+	+
Tubuliflorae	+	+
Ambrosia	+	+
Artemisia	+	+
Cruciferae	+	0
Cyperaceae	+	+
Epilobium	+	0
Galium	+	+
Gramineae	+	+
Leguminosae	+	0
Linnaea	+	0

Table 7.4 cont'd.

*Plantago lanceolata**	+	0
Plumbaginaceae*	+	0
Polygonaceae	+	0
Rumex type	+	+
Ranunculaceae	+	0
Coptis groenlandica	+	0
Thalictrum	+	0
Rosaceae	+	+
Potentilla	+	0
Sanguisorba	+	0
Umbelliferae	+	+
Spores		
Adiantum type*	+	0
*Athyrium Filix-femina**	+	0
*Botrychium**	+	0
Cysteropteris type*	+	+
*Dryopteris spinulosa**	+	+
Dryopteris thelypteris type*	+	+
Filicales undiff.	+	+
Lycopodium	+	+
L.annotinum	+	+
L.clavatum	+	+
L.complanatum type	+	+
L.lucidulum	+	+
L.obscurum	+	0
L.selago	+	0
Osmunda	+	+
Pteridium aquilinum	+	+
Selaginella	+	+
Sphagnum	+	+
Woodsia	0	+
Aquatics		
Eriocaulon	+	0
Isoetes	+	+
Myriophyllum	+	0
Nuphar	+	+
Nymphaea	+	+
Potamogeton	+	0
Typha	+	+

Key: + present 0 absent
* positive identification not possible
O macrofossil found

Taxonomic nomenclature follows Gleason and Cronquist 1963.

are exclusive to this zone). However, the percentages of herbaceous taxa present a false impression of abundance as total pollen concentrations are exceedingly low throughout the zone, ranging from 540 to 3310 grns$\cdot$cm^{-3}. Using an extrapolated sedimentation rate of 0.257mm$\cdot$yr^{-1}, total pollen influx is also meagre, ranging from 3 to 85 grns$\cdot$cm^{-2}yr^{-1}. As a result, the few arboreal pollen grains (including *Betula*) of distant origin, with the exception of *Pinus*, register small, basal peaks in the percentage diagram.

The initial rise in the percentage of *Betula* marks the transition to zone CP-2. The upper boundary also coincides with a change in the sediment stratigraphy from clay to clay-gyttja.

<u>Pollen assemblage zone CP-2</u>: *Betula*-shrubs (545-445cm; ca. 11,800-8500 BP).

This pollen zone is divided into two subzones: CP-2i, *Betula*-Cyperaceae-shrubs (545-500cm; ca. 11,800-10,000 BP); CP-2ii, *Betula-Picea* (500-445cm; ca. 10,000-8500 BP).

<u>Subzone CP-2i</u>: Steadily decreasing herb pollen percentages (32%-9%) and a relatively high contribution of shrub pollen (25%-41%) characterize this subzone. Cyperaceae peaks early (at 24%). *Myrica* maintains high percentages (7%-11%) throughout the subzone as does *Salix* (3%-6%). *Betula* increases from 21% at the base to 54% near the transition to subzone CP-2ii. Grains assigned to the Coryloid type exhibit a fairly high percentage and are probably degraded *Betula* or *Myrica*. *Populus* is consistently present as are Ericales, *Juniperus*, *Lycopodium* and Filicales. The only *Shepherdia canadensis* grains found occur in the two subzones of CP-2. The highest frequency of *Pediastrum* occurs in this

subzone, decreasing from 56% at the base to 32% at the upper boundary. In this subzone total pollen concentration ranges from less than 12,000 grns·cm^{-3} near the base to 66,000 grns·cm^{-3} at the top. Total pollen influx rises steadily from less than 300 grns·cm^{-2}·yr^{-1} at the base to 1700 grns·cm^{-2}·yr^{-1} at the upper boundary, primarily as a result of *Betula* pollen input.

Subzone CP-2ii: The transition from subzone CP2-i to CP2-ii is not well-defined (CONSLINK did not exhibit a significant split at this level) (see Figure 7.4) and is based principally on the consistent presence of, and rise in *Picea* pollen percentages (to 20%) and influx (to approximately 1100 grns·cm^{-2}·yr^{-1}) by 8900 BP. Numerical/discriminant analysis identified the presence of both *P. glauca* and *P. mariana* grains (see Table 7.5 & Figure 7.5). In addition, four *Picea* needles were discovered near the top of the clay-gyttja layer in the replicate core (see Table 7.2) and were probably deposited ca. 8600 BP.

Abies balsamea also makes its first consistent appearance in this zone after ca. 9500 BP ($\bar{x} = 4\%$). The frequencies of both *Salix* and *Juniperus* decline and the initial peak in *Betula* frequency (at 63%), corresponding to its early influx peak (ca. 2700 grns·cm^{-2}·yr^{-1}), is attained ca.9500 BP. Certain taxa, with maximal percentages in subzone CP-2i, reach their highest influx including *Salix* (117 grns·cm^{-2}·yr^{-1}), Gramineae (106 grns·cm^{-2}·yr^{-1}), and *Lycopodium* (mainly *L. annotinum* and *L. clavatum*), whereas *Myrica*, Filicales and Cyperaceae attain their initial peak in influx. *Pediastrum* declines steadily upward through this subzone. The pollen of *Quercus* and *Corylus* make their first appearance in the upper half. Minor taxa, namely *Thalictrum*, Ranunculaceae and *Sanguisorba* are present for the first time in zone CP-2. Charcoal is present in one third of the

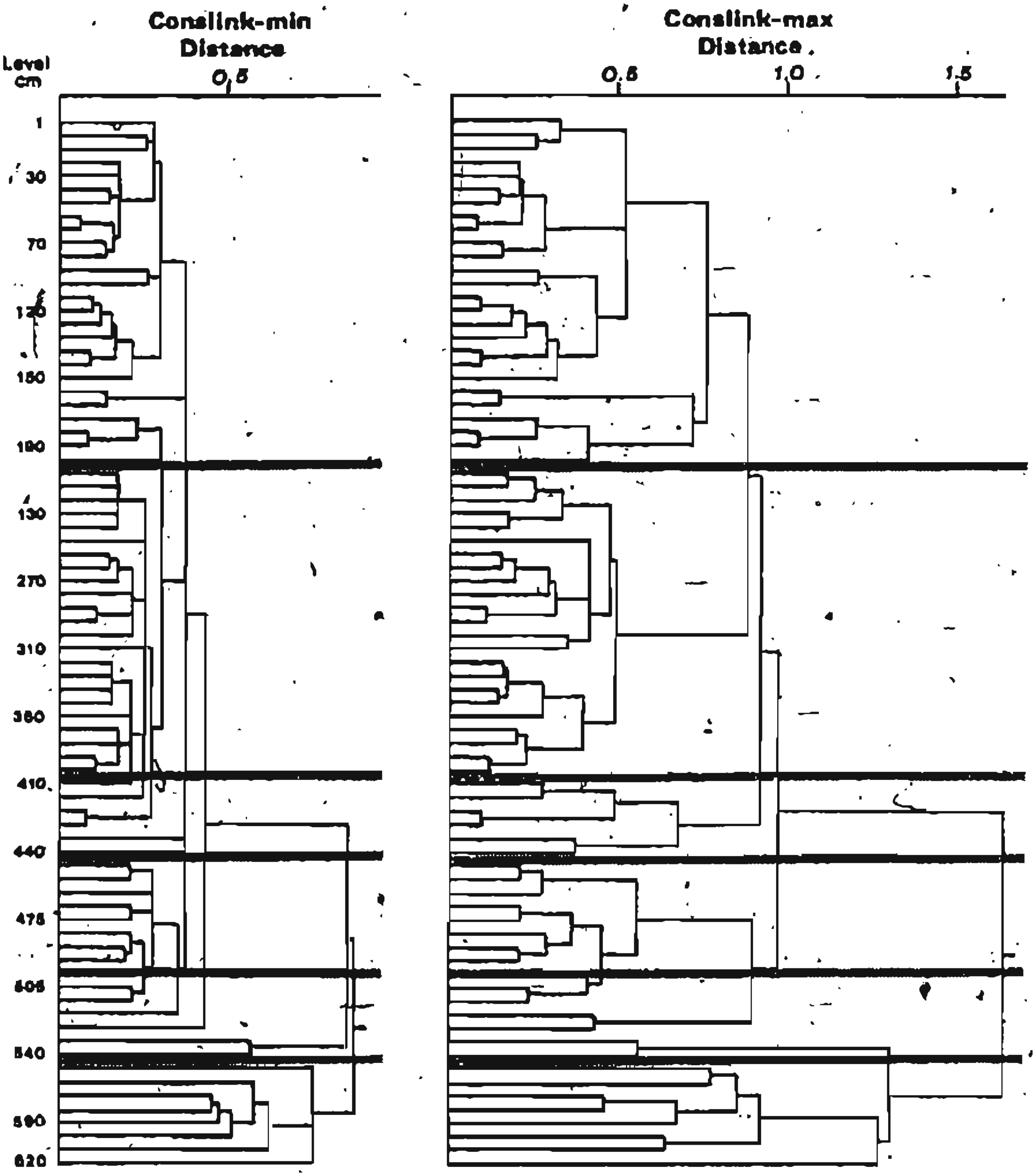

Figure 7.4: Compass Pond II, Results of CONSLINK zonation program.

Table 7.5 *Picea* Grain Determinations, Compass Pond II

	%*P.mariana*	%*P.glauca*	%unassigned
PAZ CP-4			
1cm	100	0	0
PAZ CP-3ii			
360cm	90	5	5
400cm	80	15	5
PAZ CP-3i			
430cm	70	0	30
PAZ CP-2			
470cm	45	45	10
500cm	50	50	0

n = 20 for each sample

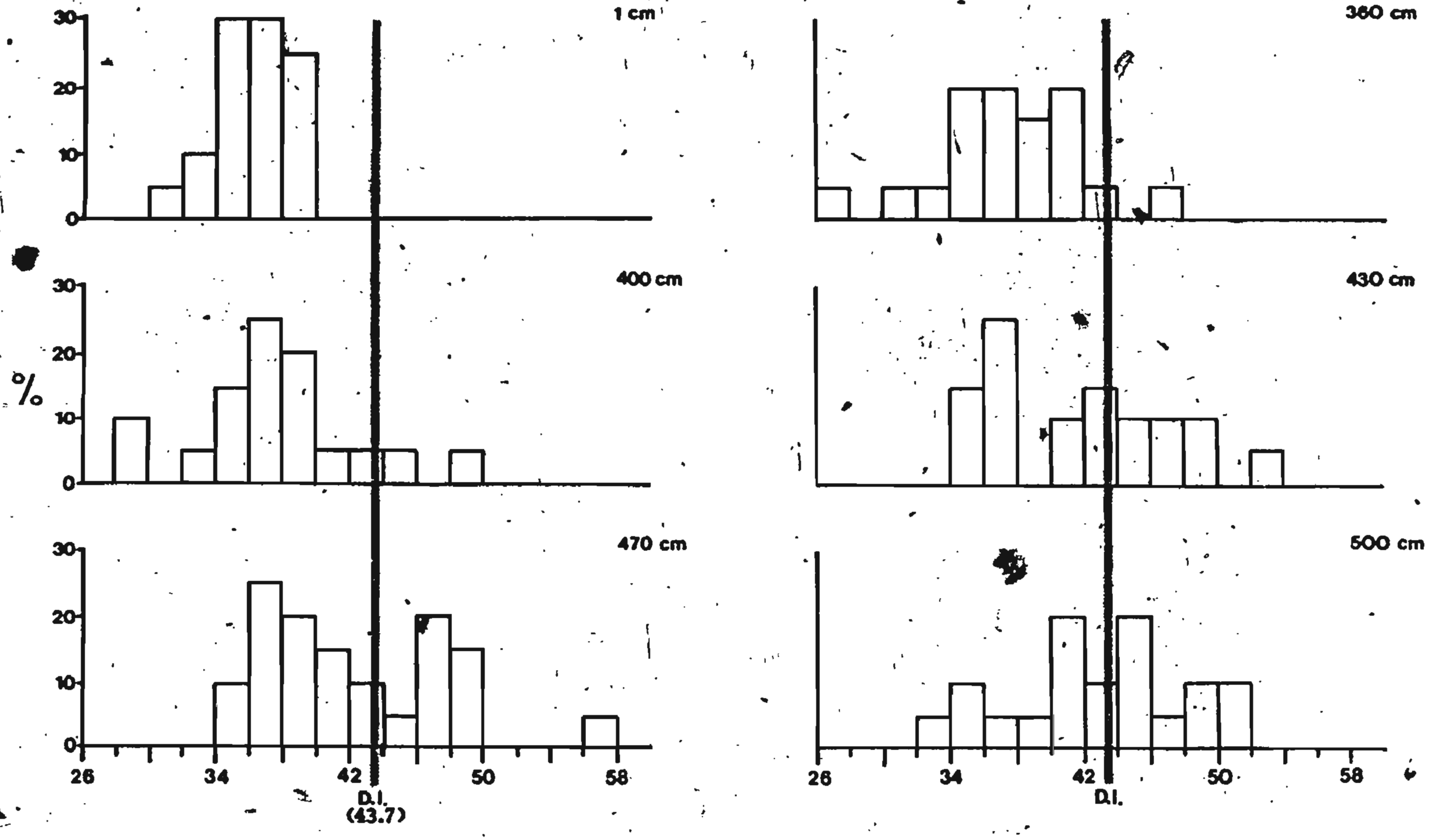

Figure 7.5: Position of fossil Picea grains on the linear discriminant function, Compass Pond II.

samples of zone CP-2. Total pollen concentration increases to over 140,000 grns·cm^{-3} and total pollen influx registers an increase from less than 2000 grns·cm^{-2}·yr^{-1} to ca. 5800 grns·cm^{-2}·yr^{-1} upward through this subzone. Increasing *Alnus* frequencies (>15%) mark the transition to zone CP-3 which is also near the change from clay-gyttja to gyttja.

<u>Pollen assemblage zone CP-3</u>: *Betula-Picea-Alnus* (445-205cm; ca. 8500-3200 BP).

The zone is characterized by the appearance of all the major constituents of the boreal forest in Newfoundland. The zone is divided as follows: CP-3i, *Alnus-Abies* subzone (445-405cm; ca. 8500-6700 BP); CP-3ii, *Betula-Picea-Alnus* subzone (405-205cm; ca. 6700-3200 BP).

<u>Subzone CP-3i</u>: An early peak in the frequency (48%) and influx (2800 grns·cm^{-2}·yr^{-1}) of *Alnus* ca. 8200 BP is followed approximately 250 years later by the maximum percentage (13%) and initial maximum influx (490 grns·cm^{-2}·yr^{-1}) of *Abies balsamea*. Near the base of the subzone and continuing upwards, an increase in the frequency of 'tree' pollen (44%-84%) depresses the shrub pollen frequencies (40%-15%). This increase in arboreal pollen appears to be primarily a result of increasing 'tree' *Betula*. (Table 7.6 shows that whereas only 7% of the birch pollen measured can be unequivocally designated as 'tree' birch at 430cm, there is a substantial reduction in the proportion of 'shrub' birch pollen at, and above, this depth.) Following a mid-zone low, both the proportion and absolute abundance of *Betula* rise. The influx curve for *Picea* shows a similar trend to the taxon's percentage curve, confirming its general decline. The pollen of certain temperate forest trees, namely *Fraxinus*, *Ulmus* and *Tsuga* together with *Taxus*

Table 7.6 *Betula* Grain Size Determinations, Compass Pond II

	x̄μm	%<20μm (shrub)	%20-26μm (indeterminate)	%>26μm (tree)
PAZ CP-4				
50cm	22.1	17	80	3
100cm	22.9	3	77	20
150cm	21.1	27	73	0
200cm	22.3	20	70	10
~~PAZ CP-3ii~~				
250cm	22.5	20	77	2
300cm	22.0	3	90	7
350cm	21.5	23	77	0
380cm	22.3	3	90	7
400cm	20.6	40	60	0
PAZ CP-3i				
430cm	22.0	23	70	7
PAZ CP-2ii				
480cm	18.6	73	27	0
PAZ CP-2i				
520cm	18.8	73	27	0

n = 30 for each sample

Betula grains measured in polar view

make their initial appearance. Other shrubs and *Populus* occur sporadically. Also, *Filicales* and *Lycopodium* are less frequent. Total herb pollen percentages are very low ($\bar{x} = 1\%$).

Total pollen influx shows an overall increase compared to pollen zone CP-2 (3700 to 6400 $\text{grns} \cdot \text{cm}^{-2} \cdot \text{yr}^{-1}$; $\bar{x} = 5000$ $\text{grns} \cdot \text{cm}^{-2} \cdot \text{yr}^{-1}$) yet records a mid-zone decline. The total pollen concentration curve exhibits a similar pattern. That peak concentration occurs at the upper boundary of subzone CP-3i is probably an artifact of the low sedimentation rate: sediment accumulated twice as fast in the preceding pollen assemblage zone, and three times as fast in the succeeding zone.

Subzone CP-3ii: The transition from subzone CP-3i to CP-3ii is not distinct but is based on a second, slower rise in the frequency of *Picea*, and a percentage peak in *Betula*. The *Picea* percentage curve lends itself to two interpretations. Either this second rise represents a subsequent increase in *P. mariana*, or the curve represents a steady rise in *P. mariana* with a superimposed early peak of *P. glauca* (see Figure A.1 in which *P. glauca* percentages are plotted on the *Picea* curve). *Betula* dominates the spectra of this subzone (37-68%) and whereas its frequency curve is uneventful, the influx curve exhibits two peak periods, namely ca. 6200-5000 BP (maximum of 6900 $\text{grns} \cdot \text{cm}^{-2} \cdot \text{yr}^{-1}$) and ca. 4700-4100 BP (maximum of 6200 $\text{grns} \cdot \text{cm}^{-2} \cdot \text{yr}^{-1}$). The percentage contribution of *Picea* is not high (12-39%), and with the exception of a slight increase near the upper boundary, the curve is complacent. The influx curve displays a similar pattern but amplifies the peaks. The same applies to *Alnus* and *Abies*: contributions range from 7-28%, and from 2-9%, respectively, and the peaks are mirrored and magnified by the influx curves.

Several other aspects differentiate this subzone from the preceding one. The pollen of *Pinus*, of distant origin, shows little proportional change but exhibits higher influx values, particularly during the period ca. 6000-3700 BP. Although the pollen of both *P. strobus* and *P. resinosa* are recorded, it was not possible to determine which species predominates as many grains could not be differentiated. Comments on certain minor taxa are worth noting. *Acer spicatum* is confined to this subzone, *Fagus* makes its first appearance in the upper half, and *Salix* virtually disappears. *Taxus* and *Fraxinus* are most frequent here and *Filicales* displays several small influx peaks. The richest assemblage of Pteridophytes is also found here. Rosaceae occurs sporadically and *Isoetes* and other aquatics are more frequent, particularly towards to top. The only grains of *Nemopanthus* and *Cornus* (*stolonifera* type) occur here. Nearly three-quarters of the samples analyzed in zone CP-3 contain charcoal.

Total pollen influx ranges from 4500 to ca. 11,000 grns·cm^{-2}·yr^{-1} ($\bar{x} = 8400$ grns·cm^{-2}·yr^{-1}). The top of this zone is defined by increasing proportions of *Picea* and *Alnus* at the expense of *Betula*.

<u>Pollen assemblage zone CP-4</u>: *Picea-Betula-Alnus* (205-0cm; ca. 3200 BP to present).

Both *Picea* and *Alnus* attain their highest percentages in this zone, up to 41% and 49%, respectively. Similarly, the influx curves for both taxa exhibit three peaks, with a maximum of 5160 grns·cm^{-2}·yr^{-1} for *Picea* and 7410 grns·cm^{-2}·yr^{-1} for *Alnus*. The first two peaks of *Picea* occur approximately 100 years after the first two peaks of *Alnus*. The percentage curves for both *Betula* and *Abies* are uneventful.

The most prominent feature of this zone occurs in the influx diagram: the lower half records the maximum total pollen influx ($\bar{x} = 10{,}600$ grns·cm^{-2}yr^{-1}) followed by a dramatic decline in the influx of *Betula*, *Picea* and *Alnus* after ca. 2000 BP. However, this striking reduction in influx does not appear to be the result of modern increased sedimentation rate. Whereas a decrease in sediment focussing has been invoked to explain a late Holocene reduction of total pollen influx to lake sites in southeastern Labrador (Engstrom & Hansen 1985), in this study, there is a correlative change in the concentration curves which, according to Pennington (1979), suggests real changes in pollen input to the lake.

Changes in minor taxa are apparent. *Myrica* exhibits three influx peaks in the lower half whereas Ericales and *Juniperus* attain their former frequency in the upper half of the zone. Aquatic species are most frequent here and herbs occur more often than in the preceding zone. In addition, *Tilia* and *Acer* (other than *A. spicatum*) are confined to this zone. Charcoal is present in every sample.

7.2 Small Scrape Pond - Results

[i] Stratigraphy and Loss-on-ignition

Table 7.7 describes the sediment of the two Small Scrape Pond cores. In contrast to the Compass Pond cores, only 9cm of mineral material was penetrated in core SSPII and none was retrieved from the other core (SSPI).

The clay-gyttja unit is 25cm (295-320cm) and 22cm (265-287cm) thick in the SSPI and SSPII cores, respectively. The organic content rises from 16.8% near the base to 25.3% near the top of the clay-gyttja unit of SSPI (see Figure 7.6). The overlying dark brown gyttja has, as expected, a higher organic content ranging from 23.4% (near the lower boundary) to 61.4% ($\bar{x} = 48.4\%$). Unlike the

Table 7.7 Stratigraphy of Small Scrape Pond Cores

Core SSPII

Depth(cm)	Sediment type	Comments
0-265	dark brown gyttja	upper 2m very wet
265-287	dark olive grey clay-gyttja	
287-296	light greenish grey clay	

Core SSPI

Depth(cm)	Sediment type	Comments
0-296	dark brown gyttja	upper 2m very wet. Plant fibres present. Spruce needles found at 240, 255 and 275cm.
295-320	medium dark brownish grey clay-gyttja	jelly consistency

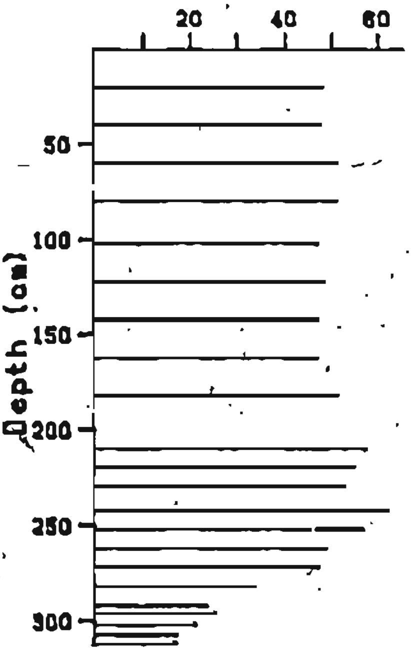

Figure 7.6: Small Scrape Pond I, Percentage loss-on-ignition.

loss-on-ignition curve for CPI, the curve for SSPI shows maximum percentages of organic matter occurring in the basal third of the gyttja unit. The bulge coincides, to a small degree, with macrofossil finds (i.e. spruce needles) in adjacent levels.

[ii] Radiocarbon dates

Radiocarbon dates were obtained for samples from the basal organic material of SSPII and the base of SSPI (Table 7.8). As neither core had more than one date, a reliable time-depth curve could not be established, but the mean sedimentation rate was 0.27 - 0.33mm yr^{-1}, approximately two-thirds of the mean sedimentation rate of Compass Pond II (0.46mm yr^{-1}).

[iii] Pollen zonation

The small Scrape Pond pollen diagrams (Figures A.4, A.5, & 7.7) are divided into local pollen assemblage zones which closely correspond to, but are not based on, changes in the sediment stratigraphy. Results of the CONSLINK program are shown in Figure 7.8.

Pollen assemblage zone SSP-1: *Betula*-Cyperaceae-shrubs (below 265cm).

This zone is divided into subzones as follows: SSP-1i; *Betula*-Cyperaceae (below 284cm; ended ca. 10,400 BP); SSP-1ii, *Betula*-shrubs-*Picea* (284-265cm; began ca. 10,400 BP).

Subzone SSP-1i: A simultaneous peak in the frequency (27%) and concentration (17,000 grns cm^{-3}) of Cyperaceae and a high proportion of *Betula* ($\bar{x}$ = 30%) characterize this subzone. The majority of taxa that have peak percentages here, namely *Myrica* (9%), Ericales (2%), *Juniperus* (6%) and Filicales (7%), all attain peak concentrations elsewhere. Other than Cyperaceae,

Table 7.8 Radiocarbon Dates for Small Scrape Pond Cores

Depth below Sediment Surface(cm)	Laboratory No.	Uncorrected Age	$\delta^{13}C$	Corrected Age	Dry Sample Wt.(g)
<u>Core SSPII</u>					
282-287	GSC-3966	10300±160	-21.7	10400±160	10.3
<u>Core SSPI</u>.					
315-320	GSC-3937	9470±160	-22.5	9520±160	12.2

Pretreatment of samples: NaOH omitted, HCl (no reaction), & distilled H_2O. Dates were based on one 3-day count in the 2L counter.

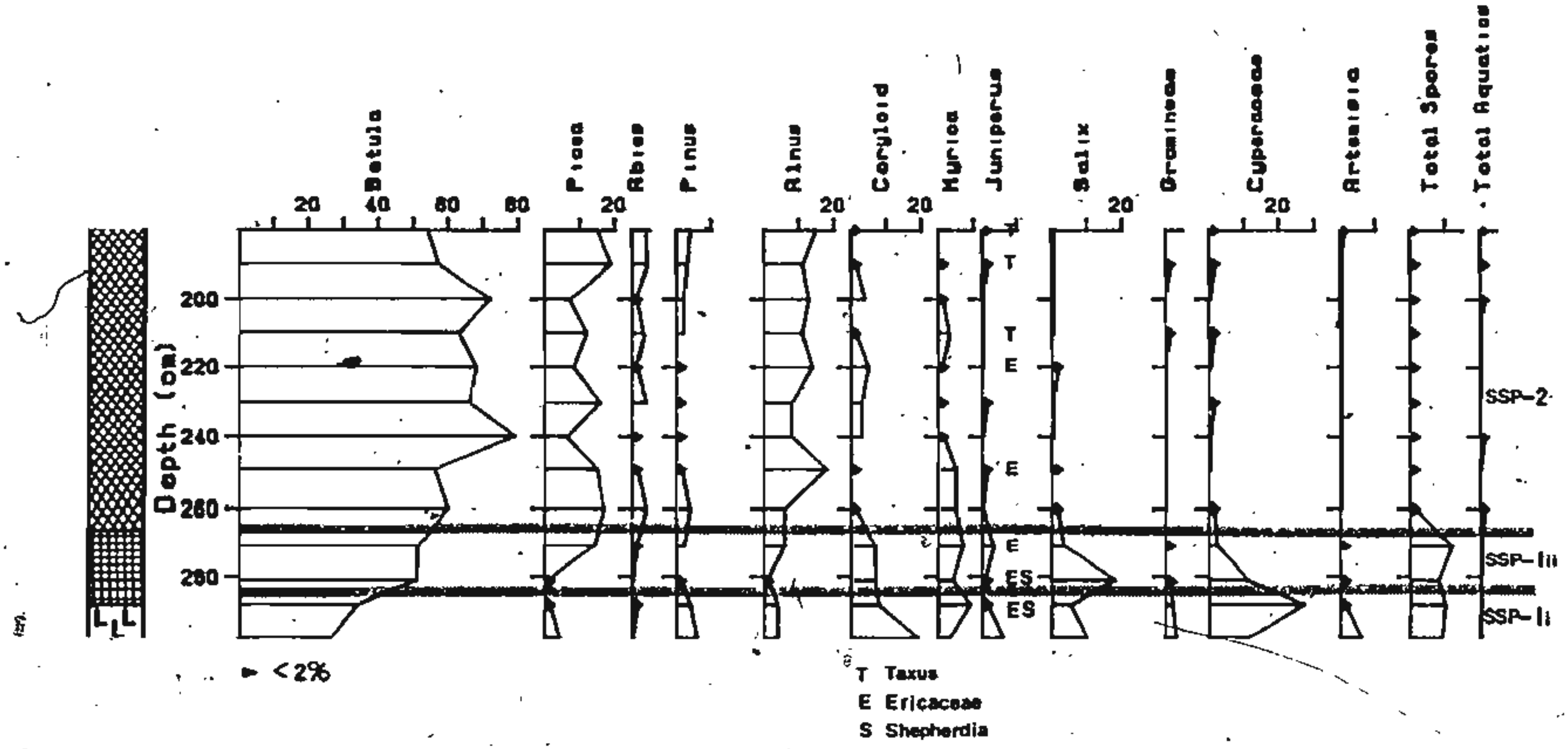

Figure 7.7: Small Scrape Pond II, Summary pollen percentage diagram.

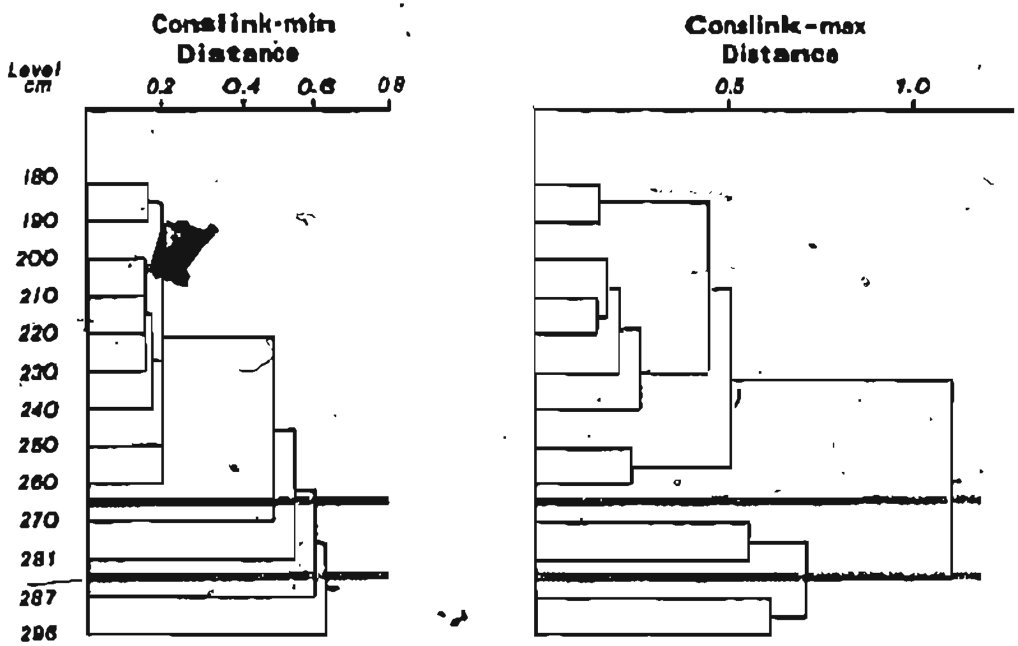

Figure 7.8: Small Scrape Pond II, Results of CONSLINK zonation program.

only *Artemisia* and *Pediastrum* exhibit both maximum frequencies and concentrations in SSP-1i. The first grain of temperate forest tree taxa, namely *Ulmus*, is found here. Total pollen concentration increases from 9000 to over 63,000 grns·cm^{-3}. The upper boundary is based on a decrease in Cyperaceae and an increase in *Salix*.

Subzone SSP-1ii: Successive peaks in the concentration of shrub taxa characterize this subzone: *Salix* (18%, 5900 grns·cm^{-3}), and *Shepherdia canadensis* (2%, 1700 grns·cm^{-3}) are followed by Ericales (2700 grns·cm^{-3}) and *Juniperus* (5400 grns·cm^{-3}). High concentrations of *Myrica* are maintained but reach a maximum in the lower part of SSP-2. The *Betula* profile of SSP-1ii is dominated by pollen from shrub sources (see Table 7.9). *Picea* exhibits a sharp rise in this subzone. Filicales is consistently present and *Lycopodium* registers a maximum in both its percentage and concentration contributions. *Populus* and *Quercus* make their first appearance. Total pollen concentration rises to over 160,000 grns·cm^{-3} at the upper boundary which is defined by an increase in the contribution of arboreal pollen and corresponds to a lithologic change from clay-gyttja to gyttja.

Pollen assemblage zone SSP-2: *Betula-Picea-Alnus* (265cm to 180cm).

In this zone the percentage curves of *Betula* (54-79%), *Picea* (6-19%), *Abies* (1-4%) and *Alnus* (6-18%) vary in methodical fashion, yet the pollen concentration curves of these taxa reveal a different pattern. In particular, *Betula* exhibits a spectacular peak exceeding 360,000 grns·cm^{-3} shortly after the initial maxima of *Alnus* (43,000 grns·cm^{-3}) and *Picea* (36,000 grns·cm^{-3}). In addition, the earliest *Picea* macrofossils were found in this zone in the replicate

Table 7.9 *Betula* Grain Size Determinations
Small Scrape Pond II

	$\bar{x}$(μm)	%<20μm (shrub)	%20-26μm (indeterminate)	%>26μm (tree)
PAZ SSP-2				
200cm	21.7	20	77	3
240cm	22.4	7	90	3
PAZ SSP-1ii				
281cm	18.6	83	17	0

n = 30 for each sample

Betula grains measured in polar view

core (see Table 7.7). Employing a mean sedimentation rate of $0.33mm \cdot yr^{-1}$, these spruce needles were probably deposited ca. 8300 BP. *Abies* pollen shows a gradual expansion from the base of SSP-2 with its highest concentration (ca. 11,000 $grns \cdot cm^{-3}$) occurring near the top of the profile (at 190cm).

The concentration of *Pinus* pollen is highest in this zone and the pollen of *Tsuga*, *Acer*, *Tilia* and *Juglans* occur sporadically. *Myrica* decreases gradually, *Salix* and Ericales are present only in the lower half and *Taxus* makes its appearance in the upper half. The pollen of aquatic plants is restricted to this zone. Charcoal is consistently present in samples above 230cm.

7.3 Comparison of Local Pollen Assemblage Zones

The earliest pollen assemblage zone of Compass Pond with high percentages of Gramineae and herbs is absent from Small Scrape Pond. Rather, the SSP-1i zone is similar in composition and total pollen concentration (9000 to over 63,000 $grns \cdot cm^{-3}$) to the CP-2i subzone (12,000 to 66,000 $grns \cdot cm^{-3}$).

In SSP-1ii, birch pollen appears to have derived primarily from shrub sources (mean size = 18.6μm), 83% measure less than 20μm) as is the case in CP-2ii (mean size = 18.6μm, 73% measure less than 20μm). However, the initial *Betula* peak registered in the Compass Pond pollen percentage diagram is absent from the Small Scrape Pond percentage diagram and both the proportion and concentration of *Betula* increase steadily.

Pollen assemblage zone SSP-2 is comparable to zone CP-3 only in that all the major components of the boreal forest are consistently present. The birch peak of SSP-2 may be similar to the second peak in the Compass Pond II profile (in CP-3ii), representing an increase in the contribution of 'tree' birch. This interpretation is supported by grain measurements which show an increase in the mean size of birch pollen in zone SSP-2 (see Table 7.9).

There is no equivalent of the Compass Pond *Alnus-Abies* subzone at the Small Scrape Pond site. In addition, *Abies* values are approximately half those recorded in CP-3i.

Based on the zonations of the Compass Pond and Small Scrape Pond pollen profiles discussed in this chapter, a tentative regional pollen zonation for the northern Baie Verte Peninsula is presented in Figure 7.9.

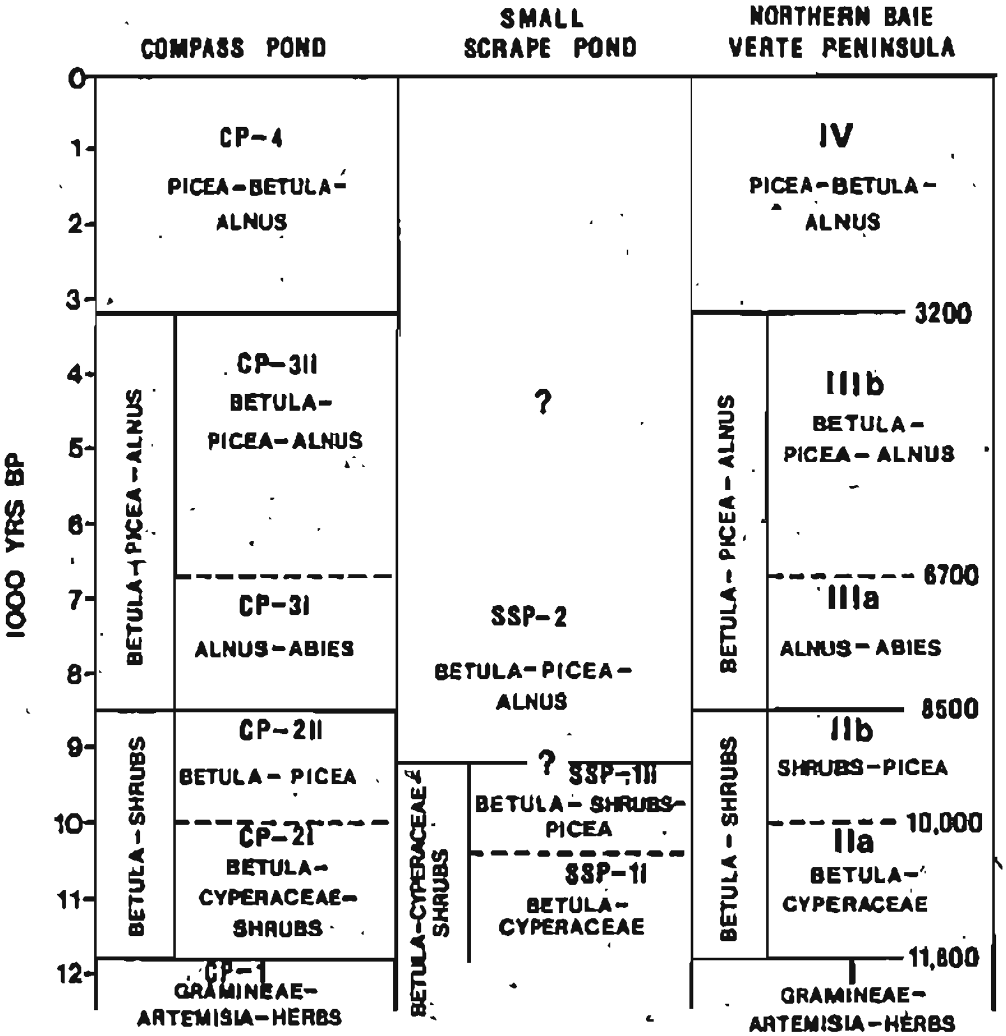

Figure 7.9: Local and proposed regional pollen zones: northern Baie Verte Peninsula.

CHAPTER 8 PALYNOLOGICAL AND STRATIGRAPHICAL INTERPRETATION

8.1 Late Wisconsin Ice Limits

Did the limit of Late Wisconsin glacial ice bisect the Baie Verte Peninsula leaving the terminus ice-free, thus providing a possible refuge for hardy plant species? Or was the area, like most of the island, ice-covered during this glaciation? Distinguishing glaciated areas from periglacial areas is difficult. For example, although the herbaceous vegetation represented by the basal pollen zone may indicate colonization of newly deglaciated ground, it may, alternatively, indicate a long-established tundra vegetation "kept open with immature soils by climatic severity" (Birks 1984, p.390). Neither do the types of plants occurring in the early assemblages always provide a clear picture of the palaeoenvironment in terms of glacial or periglacial conditions. For example, Watts (1979, p.435) suggests that "high sedge percentages are especially characteristic of recently deglaciated terrain with large areas of impeded drainage" whereas "grasses may be more characteristic of better drained or climatically dry regions" as was the case with his Longswamp site, outside the glacial limit. He also mentions, however, that high grass percentages were reported from northern Labrador in which the taxon formed part of a pioneer assemblage after deglaciation (Short & Nichols 1977).

The results of this study, in balance though not definitively, support the hypothesis of deglaciation following Late Wisconsin glaciation. The chronostratigraphic results provide the strongest evidence in support of this hypothesis. The CPII basal date corresponds with other basal terrestrial dates, in particular, that of 11,800±200 BP (GSC-3957; Macpherson in prep.) from the

King's Point site in southeastern Baie Verte Peninsula, and also accords with dated marine material ranging from 11,400 to 12,000 BP in the Baie Verte Peninsula and Notre Dame Bay areas (see Figure 8.1 & Table 8.1). Had the area been ice-free, it is possible that severe conditions led to an amictic, or perennially frozen, lake thereby delaying the onset of organic sedimentation (cf. Anderson 1983).

Sediments of cores situated beyond the Late Wisconsin ice limit may either exhibit organic material overlying till deposits (cf. Anderson 1983), or exhibit an alternating sequence of mineral and organic material corresponding to stadials and interstadials (cf. Cwynar 1982). In contrast, the cores obtained in this study show the expected sedimentary sequence for areas glaciated during the Late Wisconsin, that is basal clays grading into clay-gyttja and then gyttja (cf. Macpherson 1982b; Engstrom & Hansen 1985; Hyvärinen 1985). (The exception is the SSPI core in which there is no basal clay layer.) However, it is not possible, from the lithology alone, to unequivocally determine the depositional environment of the basal clays. It is possible that if the CP site lay within a periglacial area, then the basal mineral sediment may represent solifluction material.

The existence of laminated basal clays is often invoked as evidence that a lake received glacial meltwater (cf. Macpherson 1981). The absence of laminations in the basal CP cores may, on the other hand, reflect turbid conditions resulting from considerable slope erosion due to abundant glacial meltwater. This reconstruction is partially supported by the pattern of *Pediastrum* occurrence in the CPII profile. *Pediastrum* is generally unable to flourish if lake water is turbid during the growing season. The alga is often

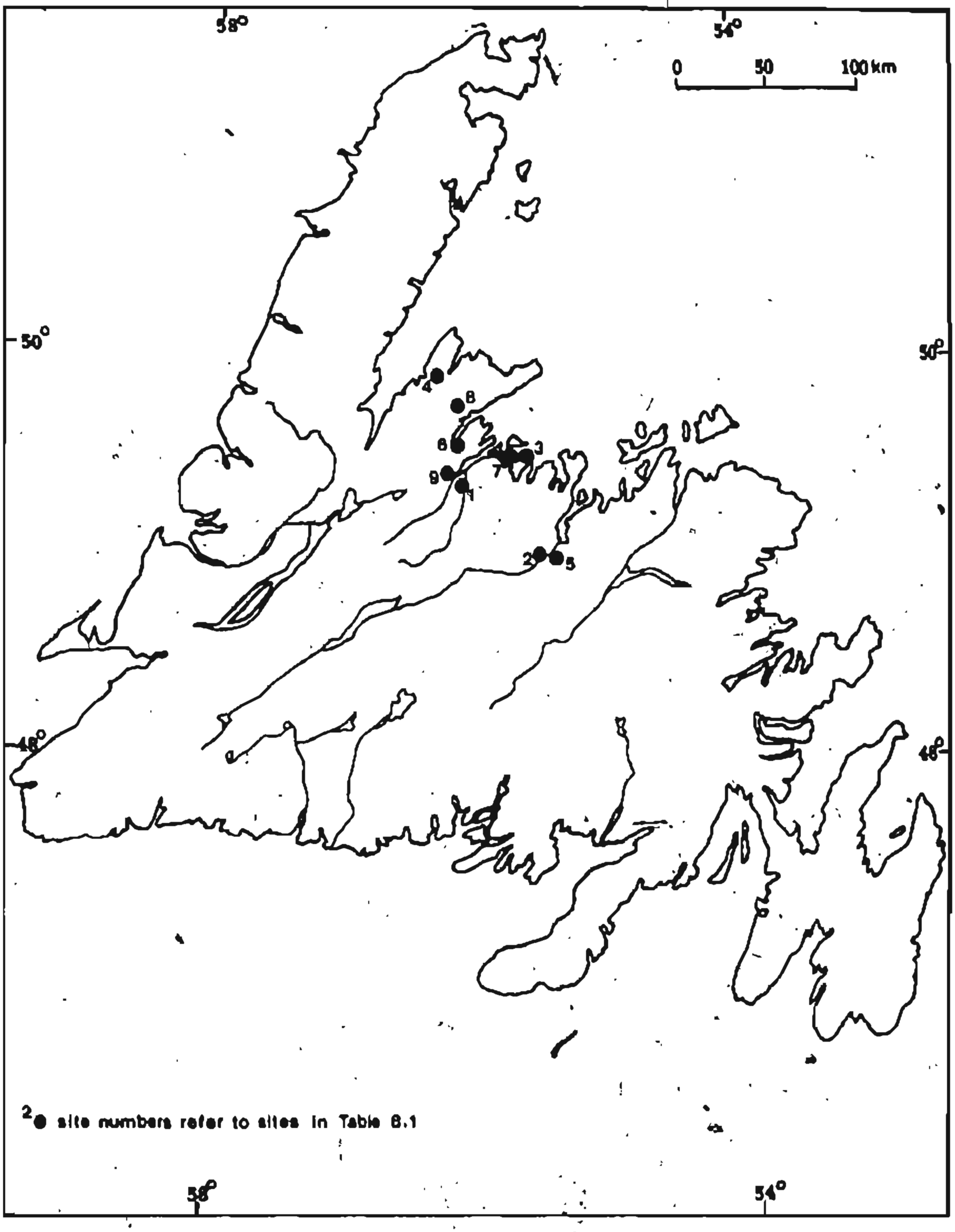

Figure 8.1: Location map of radiocarbon-dated marine material, northcentral Newfoundland.

Table 8.1 ^{14}C Dates on Marine Material near Study Area

Site & Lab.No.	Date (Yr BP)	Location & Elevation(asl)	Reference & Collector	Material & Significance
1.GSC-2085	11000±190	South Brook 49°26'N 56°04'W 4.5m	GSC Paper 75-5 C.M.Tucker	Marine clay Fixes age of glaciomarine deltas
2.GSC-3687	11400±100	Bishop Falls 49°01'N 55°28'W 3±1m	GSC Paper 83-7 D.G.Vanderveer	Shells Min. date for marine incursion
3.GSC-2318	11500±220	Triton Island 49°31'N 55°37'W 0.55m	GSC Paper 83-7 C.M.Tucker	Shell May date glacial re-advance
4.GSC-55	11520±180	Baie Verte River 49°54'N 56°17'W 48.76m	GSC Paper 63-21 E.P.Henderson	Shells Min. date for marine submergence
5.GSC-2134	11600±210	Exploits River 49°03'N 55°22'W 12m	GSC Paper 83-7 D.R.Grant	Shells Min. date for deglaciation
6.GSC-87	11880±190	Southwest Arm Green Bay 49°35'N 56°12W 10.36m	GSC Paper 63-21 E.P.Henderson	Shells Represents limit of marine submergence
7.GSC-1505	11900±200	Pilley's Island 49°30'N 55°43'W 27m	GSC Paper 83-7 N.O'Donnell	Shell Relates to late-glacial ice recession
8,GSC-75	11950±190	Middle Arm Green Bay 49°42'N 56°06'W 12.8m	GSC Paper 63-21 E.P.Henderson	Shells Relates to initial influx of sea during glacial retreat
9.GSC-1733	12000±220	South Brook 49°25'N 56°06'W 20m	GSC Paper83-7 D.R.Grant	Shell Dates postglacial marine overlap

absent from late-glacial sediments as a result of large influx of meltwater. Planktonic blooms, especially of *Pediastrum*, occur in early Holocene sediments at which time the slopes surrounding the basin are stabilized by colonizing plants yet there is still an adequate supply of nutrients to the lake (cf. Lamb 1980). The *Pediastrum* percentage curve of CPII reflects this situation: apart from an anomalously high value in the lowermost sample, the alga is absent from most of the mineral material, increases and peaks in the clay-gyttja and declines to low values in the gyttja.

The evidence taken together, including the correspondence of the CPII basal date with other dated material indicative of deglaciation or marine transgression, and the unidirectional change in the lithology coinciding with the upward increase in pollen abundance strongly implies that sedimentation was prevented by the presence of glacial ice.

8.2 Timing and Mode of Deglaciation

If it is assumed that the study area lay within the Late Wisconsin ice limit, the next question to be addressed concerns the timing and mode of deglaciation. Estimating the time of deglaciation can only be achieved indirectly by dating the basal layers of organic sediments. Again, this will produce only a minimum date of deglaciation because (a) it is not known how long after deglaciation the accumulation of organic matter began, and (b) the date obtained does not relate to the very base but is an average[1] age of the entire sample thickness. Thus the true date of initial organic deposition will be older than the date ascribed to a

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[1] the radiocarbon date is not really an average of the sample as ^{14}C decay is exponential not linear.

relatively thick basal sample (Sutherland 1980). The degree of 'younging' will depend upon the rate of sedimentation. Sutherland (1980) suggests that at a time of soil instability and sparse vegetation cover, sedimentation rates would have been relatively fast. By contrast, Lundqvist (1980) claims that organic sedimentation close to the receding ice front would probably have been exceedingly slow.

Estimates of the period of accumulation of basal organic samples have been made for certain Scottish lakes (Sutherland 1980). Based on overall rates of sedimentation and thickness of the basal samples, Sutherland calculated the dates to be from 50 to 200 years younger than the actual base of the sample. A greater 'younging' effect was determined for the basal organic deposit of Proteus Lake, Ellesmere Island. A radiocarbon assay on the basal 0.5cm sample, using accelerator mass spectrometry, produced a radiocarbon date 400 years older than the date ascribed to the basal 5cm sample (Blake 1985).

Applying the technique of Sutherland (1980) to the basal organic samples of CPII and SSPII it is possible to obtain a closer approximation of the onset of organic sedimentation at these sites. Table 8.2 indicates that the basal dates for the two sites are approximately 100 years (half the accumulation period) younger than the actual base of the samples, which is less than the noted standard deviations. It is interesting to note that this is the figure that King (1985b) added to basal dates of lake sediments in Labrador in his evaluation of radiocarbon dates of local deglaciation. These figures thus suggest that 11,800 and 10,500 BP are minimum dates for deglaciation of the Compass Pond and Small Scrape Pond sites, respectively.

Table 8.2 Estimated Periods of Accumulation of Basal Organic Samples

Site	Rate of Sedimentation (mm·yr^{-1})	Basal ^{14}C Date (yrs BP)	Basal Sample Thickness(mm)	Period of Accum(yr)
CPII	0.257[1]	11,700±180	50	195
SSPII	0.274[2]	10,400±160	50	182

[1] rate calculated from two lowermost radiocarbon dates

[2] rate for entire core based on basal radiocarbon date

The basal date (both given and estimated) of Small Scrape Pond is 1300 years younger than the minimum date for deglaciation of Compass Pond, 15km to the northwest and ca. 114m higher in elevation, and suggests downwasting of the ice in northern Baie Verte Peninsula. This contrasts with evidence from other terrestrial sites (GSC-3608, -3646, -3634; Blake 1983) that indicates a pattern of ice recession toward the height of the land in northcentral Newfoundland.

Yet downwasting was not the sole mode of ice recession on the Baie Verte Peninsula. Whereas the King's Point site has a minimum date of deglaciation of 11,800 BP, its elevation is only ca. 102m.a.s.l. It appears, therefore, that the dissipation of Late Wisconsin ice from the area was in the form of downwasting together with ice recession toward the interior of the peninsula. Although not explicitly stated, a study by Tucker (1974) implies that this was the case. He concluded that the coastal areas near Halls Bay and Green Bay were quickly deglaciated by calving ice fronts and that ice retreated from the coast by stagnating in valleys and lowlands. For the southern part of the peninsula, Tucker proposed that "final topographic flow into the Indian Brook Valley may have been from the plateau areas to the north and south and that residual ice remained in the upper part of the valley for a considerable time after ice retreated from the coast" (1974, p.108).

The evidence from this study strongly points to the existence of a residual ice cap on the Baie Verte Peninsula as Lundqvist (1965) earlier concluded. That the area was deglaciated later than other parts of northcentral Newfoundland, the Leading Tickles site in particular, may be explicable in terms of different climatic regimes. The present climate of northern Baie Verte Peninsula is cooler, in terms of the annual average number of degree-days above 5°C, than the area south of Notre Dame Bay (Banfield 1981) and is influenced, indirectly, by the cold Labrador current and by the persistence of ice in bays until early into the summer season (Farmer 1981). In this regard it is interesting to note that deglaciation of the Fogo Island site (49°42.5'N 54°15.5'W), approximately 175km east of the CP site, was delayed until about 11,000 BP (GSC-3973; Macpherson in prep.). The higher elevation of the Baie Verte Peninsula and associated decrease in atmospheric temperature of about 1°C vis-a-vis the Leading Tickles site may also help to account for the relatively delayed deglaciation of the area.

8.3 Vegetation Reconstruction of Compass Pond Site

As mentioned in Chapter 4, a variety of approaches to vegetation reconstruction was employed including the use of modern analogues. The comparison of fossil pollen assemblages with modern pollen assemblages was done primarily using modern pollen spectra from other regions due to the absence of modern pollen-vegetation studies for the study area. It is interesting, however, to compare the proportions of individual tree taxa in the modern pollen spectrum of Compass Pond with the proportions of trees (see Table 5.3) in the vegetation of northcentral Newfoundland. Such a comparison indicates that, with the exception of spruce, the taxa are either grossly over- or underrepresented by their pollen. Spruce comprises about 50% of the modern tree pollen spectrum and 54% of the productive forest. Balsam fir makes up 27% of the forest yet contributes only 4% of the modern pollen input to the sediments of Compass Pond. In contrast, birch is strongly over-represented, comprising 44% of the modern tree pollen spectrum of Compass Pond but less than 12% of the productive forest. Although birch pollen grains were not quantitatively differentiated as to 'tree' and 'shrub' types, a visual inspection suggests that the majority of the grains are of birch 'tree' origin. Certain trees, including larch and white spruce, are not represented at all in the modern pollen spectrum and hardwoods are poorly represented by their pollen.

(i) <u>Gramineae-*Artemisia*-herb zone</u> (ended ca. 11,800 BP):

The pollen assemblage of this zone indicates that the site was receiving pollen from both extra-regional sources, primarily *Pinus*, renowned for its dispersal ability, and local, mainly herbaceous, plants. The high proportions of Gramineae and *Artemisia* have no known modern analogue from Newfoundland but similar assemblages have frequently been reported as occurring in late-glacial profiles (Anderson 1980). The most similar fossil assemblage from a Newfoundland site is that described for pollen assemblage zone LT-2, from Leading Tickles, which ended ca. 10,500 BP. There, the late-glacial spectra, interpreted as a sparse herb tundra, comprise high percentages of Gramineae and *Artemisia* (about 10%) and associated taxa of open habitats including Caryophyllaceae and Chenopodiaceae (Macpherson & Anderson 1985). In contrast to the herb zone of Compass Pond however, chionophilous taxa, *Oxyria* in particular, are present in LT-2.

Artemisia-dominated assemblages have also been well documented for the Wisconsin glacial stage in northwestern Canada and Alaska (e.g. Colinvaux 1964; Cwynar 1982; Ritchie 1985) and the Younger-Dryas stadial in northern Britain (e.g. Walker 1975; Birks & Mathewes 1978). Because *Artemisia* is regarded as an indicator of open and disturbed soils, and because these herb pollen assemblages are characterized by low pollen influx, most researchers in these areas have interpreted the Gramineae-*Artemisia* pollen association as representative of a steppe, fellfield or simply 'tundra-like' vegetation.

Finding a modern analogue for late-glacial and early Holocene pollen assemblages is a particularly thorny problem since the communities from which the pollen were derived were undoubtedly controlled by edaphic conditions and biotic factors (e.g. competition, migration) that were vastly different from present-day situations. Surface samples from the Canadian arctic and subarctic tundra, for example, record high percentages of Gramineae pollen (Ritchie & Litchi-Federovich 1967), yet *Artemisia* values generally comprise less than 4% of the pollen sum (Ritchie & Litchi-Federovich 1967; Terasmae 1967, 1976; Davis & Webb 1975) and the genus is absent altogether from high arctic vegetation (McAndrews 1984).

Perhaps the closest modern analogue for the CP-1 pollen zone is that reported by Pennington (1980) from the low-arctic continental region of western Greenland. The steppe component of this region, occupying dry areas, has been described as sub-low arctic continental xerophytic grasslands and dwarf shrub vegetation with associated willow scrubs (Böcher 1954). Gramineae (up to 77%) and *Artemisia* (2-40%) dominate the modern spectra (Pennington 1980). The Greenland form of the latter taxon is *A. campestris* ssp. *borealis*, found frequently where plant cover is discontinuous, particularly on "exposed slopes and ridges from which the protective snow cover appears to be blown" (Pennington 1980, p.177) and "its optimum is...in the most xerothermous sociations" (Böcher 1954, p.188). The same subspecies (var. *latisecta*) occurs today in Newfoundland together with two other subspecies, namely ssp. *canadensis* and ssp. *caudata* (Scoggan 1979).

The representation of *A. borealis* in modern surface samples in western Greenland is closely related to the range of the plant (Pennington 1980). It seems reasonable to assume that most of the *Artemisia* pollen recorded in the Compass Pond profile is derived from the indigenous subspecies of *A. campestris* and, furthermore, the percentage peak suggests that the taxon was an important component of the early plant communities.

Yet the difficulty in interpreting the Compass Pond herb zone remains, for although the composition and, to a certain degree, the proportions of the dominant taxa suggest an affinity with the steppe of western Greenland, the vegetation cover must have been considerably sparser. Both the total pollen concentration (540-3310 grns·cm^{-3}) and estimated influx (3-85 grns·cm^{-2}·yr^{-1}) of CP-1 are far lower than those recorded for low arctic tundra sites in either Greenland or eastern North America (cf. Davis & Webb 1975; Pennington 1980). The absolute pollen values of the herb zone are closer to those reported from rock desert and tundra areas of the Canadian high arctic (5-65 grns·cm^{-2}·yr^{-1}) (Ritchie & Litchi-Federovich 1967), or the high arctic of northern Greenland (6-8 grns·cm^{-2}·yr^{-1}) where the vegetation cover is less than 3% (Fredskild 1973).

Reconstruction of the various plant community immediately following deglaciation is difficult because it was possible to identify the pollen of local plants only to the level of the family; many of these contain species of varying ecological affinities. It appears that this herb zone represents a pioneer vegetation phase and was unlike any communities found today in high arctic or low arctic regions. The presence of both *Artemisia* and *Sphagnum* suggests a mosaic of dry and damp habitats. Unstable slopes and wind-swept ridges provided suitable conditions for the growth of *Artemisia* and other open ground herbs including members of the Caryophyllaceae, Chenopodiaceae, Cruciferae, Tubuliflorae, Leguminosae and Rosaceae. *Sphagnum* and possibly sedges were restricted to stream banks and the lake margin. Graminoids may have grown on exposed upland sites as well as near the lake's edge as *Calamagrostis* does today. The absence of chionophilous taxa (e.g. *Oxyria*, Ranunculaceae), in addition to the occurrence of *Artemisia*, frequently associated with early snow melt (Pennington 1980) and summer drought (Böcher 1952), suggest that there were few, if any, snowbed communities. While the proportions of Ericaceae and *Salix* are not high, the pollen grains of many members of these taxa are poorly adapted for long-distance dispersal implying that patches of dwarf shrubs grew in the vicinity. Later, juniper shrubs invaded the well-drained sites of the upland communities.

The discontinuous cover of the vegetation during this phase resulted in slope instability and enabled high inputs of mineral matter to enter the lake as is attested to by the low organic content of the basal sediments. Although the lake was nutrient-rich during this stage, it was almost barren of aquatic species. Again, this contrasts with the present-day situation in the Canadian arctic (Terasmae 1967) yet may be explained in terms of turbid conditions.

Of all the zones of CPII, the herb zone exhibits the least diverse flora in terms of the mean number of taxa identified per sample (i.e. 15, including extra-regional pollen). This somewhat impoverished flora may be an artifact of the low pollen sum and high percentage of unidentified grains. On the other hand, a species-poor vegetation during this period could represent a relative dearth of habitat types, particularly mesic sites. A study on plant immigration in front of retreating glaciers (Stork 1963) has shown that species-richness increases as plant cover closes and where water supply (ground or meltwater) is sufficient. The largest number of species identified (26) in any one level occurs near the top of this zone and corresponds with a change in the lithology.

(ii) Betula-shrubs zone (ca. 11,800-8500 BP):

As with the preceding zone, this fossil assemblage lacks any direct analogy with modern spectra from Newfoundland or eastern Canada. The continuous presence of *Populus* pollen and insignificant contributions of *Alnus* and *Picea* pollen in the CP-2i subzone contrast with the occurrence of these taxa in surface samples from tundra and forest-tundra areas (cf. Davis & Webb 1975; Lamb 1984). However, the *Betula*-Cyperaceae-shrub subzone exhibits an affinity with the dwarf-shrub heath tundra community described by Terasmae (1967) for northwestern Canada in terms of the dwarf birch, Ericaceae and Cyperaceae components, although there the extant vegetation does not include *Salix*, *Myrica* or *Populus*. Furthermore, open dish samplers from dwarf-shrub tundra communities in the Canadian low-arctic (Ritchie & Lichti-Federovich 1967) have a pollen content broadly corresponding to the CP-2i subzone, although *Myrica* is rarely registered and the total pollen influx (50-760 grns$\cdot$cm$^{-2}\cdot$yr^{-1}) is half that of CP-2i.

A compositional similarity between the CP-2i subzone and the LT-3 zone from northcentral Newfoundland is apparent although the succession of shrubs differs and *Populus* is absent from the latter assemblage. Macpherson and

Anderson (1985) interpreted the vegetation of LT-3 as being comparable in appearance to the low-arctic dwarf-shrub heath of coastal south Greenland which is characterized by ericaceous plants and sedges. However, a dissimilarity between these fossil assemblages and modern assemblages from Greenland is that *Myrica* is not indigenous to that country. Total pollen influx varies from a mean of 250 grns$\cdot$cm$^{-2}\cdot$yr^{-1} at the coast to a mean of 2180 grns$\cdot$cm$^{-2}\cdot$yr^{-1} near the heads of the fjords (Fredskild 1973).

Although the *Betula*-Cyperaceae-shrubs subzone may best be described as representing unique and ephemeral associations of species, it may have resembled, at least physiognomically, a low-arctic dwarf-shrub tundra. An important exception to this was the likely presence of *Populus* trees. What was the environmental setting during this period? No doubt open-soil plants (e.g. *Artemisia*, Caryophyllaceae) suffered competition from invading shrubs. Populations of ericaceous shrubs, *Myrica* and 'shrub' birch in the region increased gradually between ca. 11,800 and 10,000 BP. The greatest populations of *Salix* shrubs occurred between ca. 10,400 and 9500 BP, broadly corresponding with the increase in sedge populations. This may have represented the presence of sedge-willow communities and an increase in the extent of moist sites resulting from impeded drainage. Clubmosses and ferns also increased in abundance.

It is interesting to note that the displacement of dwarf shrubs by taller shrubs (i.e. *Myrica*, *Betula*) took a longer period at Compass Pond than it did at Leading Tickles. A logical explanation for this is that ice remained longer at lower elevations on the Baie Verte Peninsula thereby inhibiting or delaying species migration. On the other hand, it is suggested that *Populus* immigrated to the area during an earlier stage of vegetational development than it did at either Leading Tickles on the northcentral coast or in northeastern Avalon (cf. Macpherson 1982b, & in prep.). Although the proportion of *Populus* is not high during the *Betula*-shrubs zone, the taxon's frequent underrepresentation is well known. For example, in the forest-tundra south of James Bay, where both *P. tremuloides* and *P. balsamifera* form stands on cut-overs and burned areas, the regional *Populus* pollen frequency falls below 1% (Farley-Gill 1980). Similarly, stands of *P. tremuloides* occur in northcentral Newfoundland and the species is found locally near Compass Pond in unshaded sites yet the uppermost samples from CPII reveals less than 1% of *Populus* pollen (and only 5 grns·cm^{-2}yr^{-1}).

Studies by Richard (1978) and Mott and Farley-Gill (1981) in Québec have revealed that *Populus* was a forerunner in the process of afforestation, as appears to have been the case in northern Baie Verte Peninsula, and was later replaced by spruce. At their sites, the abundance of *Populus* pollen during the early vegetational stages (>30% & >7000 grns·cm^{-2}yr^{-1}, respectively) was far greater than that at Compass Pond ($\bar{x}$<3%, 3 grns·cm^{-2}yr^{-1}). However, whereas Mott and Farley-Gill (1981) suggest that *Populus* occurred in clumps or groves, after it invaded the Gatineau Park area ca. 10,700 BP, it is suggested that at Compass Pond, only isolated trees occurred on the generally open landscape.

Therefore, although the genus appears to have played a less significant role in the tundra-forest transition than it did in southwest New Brunswick (Mott 1975a) and Québec (Richard 1978; Mott & Farley-Gill 1981), nevertheless, *Populus* may have occurred locally as early as 11,000 BP. In contrast to other arboreal taxa, there would likely have been few barriers to the migration of

Populus. Its opportunistic strategies, including widely disseminated propagules and pioneering ability on mineral substrata (Page 1972; Mott 1978; Mott & Farley-Gill 1981; Ritchie 1984) make it an ideal harbinger of arboreal immigration.

The approximately 1500 year span covered by subzone CP-2ii witnessed several major floristic and structural changes in the vegetation. Also, the number of taxa identified is highest in this subzone indicating a comparatively species-rich vegetation. Dwarf shrubs became increasingly repressed as populations of shrub birch, peaking at ca. 9500 BP, and *Myrica* expanded. Spruce invaded the shrub tundra ca. 9300 BP. Balsam fir arrived about 100 years later, preceding the expansion of alder. The immigration of fir following spruce has also been reported from western (Brookes 1981), northcentral (Macpherson in prep.) and eastern Newfoundland (Macpherson 1982b) and southern Labrador (Engstrom & Hansen 1985).

Grain size determinations have shown that white and black spruce arrived simultaneously in the area as they did in central Labrador (Lamb 1984). However, as *Picea glauca* appears to produce less pollen than *P. mariana* (Lamb 1984), the early populations of *Picea* near Compass Pond were probably dominated by white spruce. Today, white spruce accounts for about 2% of the northcentral Newfoundland forest (Govt. of Nfld 1974) and it is more abundant closer to the coast (Damman 1983). Yet the *Picea* grains measured in the uppermost sample of CPII were all derived from black spruce. In other regions, from northeastern United States to southern Labrador, white spruce preceded black spruce (Watts 1979; Lamb 1980; Davis 1983; Engstrom & Hansen 1985).

Davis (1983) postulates that the neutral or calcareous soils of New England during the late-glacial may have been conducive to the growth of white spruce. Evidence that similar edaphic conditions probably persisted in the study area is provided, in part, by the sporadic occurrence of such calciphilous taxa as *Shepherdia canadensis* and *Lycopodium clavatum* (Birks 1973). White spruce may have colonized upland sites and the understory of poplar groves, later replacing the light-demanding poplar as it did in northwestern Canada (Ritchie 1985). Studies have shown that white spruce seedlings are shade-tolerant and their survival is aided by a partial canopy which tends to reduce both moisture loss and surface temperatures (Jarvis et al. 1966; Richardson & Hall 1973b). However, the vegetation near Compass Pond was unlike the forest-tundra of present-day Labrador in which alder is abundant, balsam fir is rare and *Populus* is absent (Lamb 1984).

Although trees grew in the area between 10,000 and 8500 BP, they may have been restricted primarily to favourable sites, so that the vegetation remained fairly open. Sustained influx values of *Betula* and *Myrica* in subzone CP-2ii provide evidence that the population densities of shrub birch and *Myrica* within the vegetation were maintained. In addition, lowland sites may have favoured the growth of tall-herb communities comprising *Sanguisorba*, *Rumex*, *Thalictrum* and other Ranunculaceae.

The loss-on-ignition curve reflects a decreasing minerogenic input to the lake resulting from the gradual establishment of vegetation cover and inwash of humus from developing soils. The lake flora remained depauparate during the *Betula*-shrubs zone although *Pediastrum* flourished in the early stages. The sudden peak of the alga at the base of the zone, corresponding to the change from clay to clay-gyttja, implies that the lake became decreasingly turbid. The gradual decrease of *Pediastrum* is probably a consequence of fewer nutrients entering the lake.

(iii) ***Alnus-Abies* subzone** (ca. 8500-6700 BP):

The anomalous association of *Alnus* and *Abies* in CPII has no modern analogue from Newfoundland or Labrador and may, again, represent the individualistic behaviour of species having different migration rates. Comparable fossil assemblages are absent from Newfoundland but have been reported from southern Labrador. There, Lamb (1980) designated an *Alnus-Abies* Regional Pollen Assemblage Zone as occurring between ca. 9000 and 5000 BP, although in the local assemblages of his three sites *Alnus* always preceded *Abies*. He later refers to this stage as a fir-dominated forest (Lamb 1984). Elsewhere in southeastern Labrador, Engstrom and Hansen record[illegible]ce-fir zone with an alder peak between 8000 and 6000 BP which they cla[illegible]rds "the transformation of shrub-tundra to conifer forest" (1986, p.550).

Was it a fir-dominated forest that covered the hills surrounding Compass Pond during this period, or was it an open spruce-fir woodland? It is difficult to determine the density of the vegetation cover because of the contrasting signals emitted by the absolute pollen curves. The high total pollen concentration values during this period are akin to those from surface-samples of the boreal forest (Davis & Webb 1975), suggesting that the area was completely tree-covered. In contrast, the total pollen influx, exhibiting no appreciable increase over the preceding zone, suggests that the tree canopy remained open.

Undoubtedly the *Alnus-Abies* subzone reflects a period of rapid vegetational change. Percentage loss-on-ignition continues to increase throughout this subzone and the lithology changes from clay-gyttja to gyttja. The period was characterized by (a) the steady expansion and maximum density of alder populations ca. 8180 BP followed by (b) a peak in the fir population ca. 7900 BP, which (c) coincided with an increase in black spruce at the expense of white spruce, and (d) a decrease in 'shrub' birch and the immigration of 'tree' birch into the region.

Workers in southeastern Labrador (Lamb 1980; Engstrom & Hansen 1985) point out that the magnitude of the *Alnus* peak in fossil assemblages from that area resembles the taxon's maximal percentages in surface-samples from the forest-tundra. The proportion of *Alnus* in the fossil spectra following its peak at CPII is closer to its representation in the modern boreal forest (cf. Davis & Webb 1975) and probably reflects the subsequent closing of the forest canopy. The initial expansion of alder populations may have been favoured by the presence of birch. Such an association occurs at the modern forest line in northern Québec with birch providing protection for alder shrubs against wind (Gilbert & Payette 1972). Later, an extensive cover of alder thickets would have improved soil fertility by increasing the amount of available nitrogen and organic matter. An increase in organic matter is implied by the comparatively high percent loss-on-ignition during this subzone. Speculation as to the role of *Alnus* in forest succession in Labrador led Lamb (1980) to suggest that the added nitrogen afforded by the genus may have been beneficial to the growth of *Picea*. However, the arrival of *Picea* near Compass Pond preceded that of *Alnus*, as was the case in most of eastern North America (Davis 1981).

The decrease in white spruce may have been due to the immigration and competitive superiority under optimal conditions of *Abies*, as Lamb (1980) suggests for southeastern Labrador. Changing edaphic conditions, namely an increase in organic matter and corresponding reduction of exposed mineral soils, may also have contributed to a reduction of suitable habitats for white spruce.

The maximal *Abies* percentages (>12%) of the *Alnus-Abies* subzone are higher than those of the uppermost samples of CPII (<5%). Today, *Abies* comprises almost one third of the northcentral Newfoundland boreal forest (Govt. of Newfoundland 1974) suggesting that the representation of fir in the early forest was considerably greater. Yet the canopy of this early forest may not have been dominated by fir as it appears to have been further north where *Abies* pollen reached 20-30% (Engstrom & Hansen 1985). Nearly simultaneous decreases after ca. 7600 BP in the percentage, influx and concentration of *Abies* indicate a real decline in fir populations.

A decline in *Abies* populations within the early forest cannot be easily interpreted or explained and may have been the result of several interacting factors. Lamb (1980) hypothesized that in southern Labrador this phase was a result of soil deterioration and increased fire frequency: two conditions that would give black spruce the competitive edge over fir. Engstrom & Hansen also stress competition suggesting that "marginal habitats were quickly filled by black spruce...and fir-dominated stands were increasingly confined to sites [with] good soils" (1985, p.551).

Unlike the pollen diagrams from southern Labrador, the CPII diagrams exhibit several subsequent *Abies* peaks suggesting a resurgence of fir in the boreal forest stage. Had soil deterioration been the main cause of the taxon's initial decrease, it would be difficult to account for a later fir increase as a reversal in edaphic conditions. The evidence here suggests that an increase in fire frequency was an important factor in decreasing fir populations. It appears that after ca. 8500 BP an open forest or woodland, with a canopy comprising fir and spruce, developed in the region. Since *Alnus crispa* today grows in a variety of sites in northern Baie Verte Peninsula (see Chapter 5), the presence of alder during this zone may indicate a mosaic of vegetation types. Increased litter accumulation and drier conditions may have facilitated an increase in outbreaks of fire. Such disturbances would have favoured the regeneration of black spruce which, unlike white spruce or fir, maintains a fairly constant seed population which is not destroyed by fire (Rowe & Scotter 1973; Black & Bliss 1980) and is well-adapted to immediate post-fire reproduction. An increase in the presence of charcoal in the CPI samples and the occurrence of *Epilobium*, albeit a single grain, support the hypothesis that fire has always played an integral part in the forest history of this region.

(iv) _*Betula-Picea-Alnus* subzone_ (ca. 6700-3200 BP):

By 6000 BP all the major constituents of the modern forest were in place and the transition from an open to a closed forest cover was complete. Pollen percentages during this subzone indicate that a comparatively stable period ensued for approximately 3000 years. Total pollen concentrations (74,000-255,000 grns$\cdot$cm^{-3}) of the CP-3ii subzone imply boreal forest conditions (cf. Davis & Webb 1975; Macpherson 1982b) and total pollen influx (4500-11,000 grns$\cdot$cm$^{-2}\cdot$yr^{-1}) is similar to that of the spruce zone of southern Labrador, interpreted as a closed-crowned forest (Engstrom & Hansen 1985).

However, the composition of the forest did not replicate the present vegetation formation nor was it similar to the early boreal forest of southern Labrador. The most striking difference was the lower proportion of spruce pollen at the Compass Pond site caused by the greater deposition of 'tree' birch pollen. Although the majority of these birch grains fall within the size range representing either shrub or tree origin, the second birch rise in the diagrams is ascribed to the immigration and expansion of 'tree' birch, as it has been in other regions (e.g. Lamb 1980; McAndrews in Brookes 1981; Macpherson 1982b) (see section 8.6).

Since *Betula papyrifera* is common today in the Baie Verte Peninsula, it appears reasonable to assume that it was this species that coexisted with black spruce 6000 years ago. As an increase in both white birch and black spruce is compatible with increased fire frequency, it is argued here that a white birch-black spruce fire forest existed during this subzone and that increased fires were due to warmer and probably drier conditions.

The primary evidence of a climatic optimum is the small but significant rise in the influx, and to a lesser degree the concentration, of *Pinus* pollen between ca. 6000 and 3700 BP. That this was a regional and not local event is attested to by the contemporaneous *Pinus* maximum registered at the Leading Tickles site (Macpherson in prep.). Neither white nor red pine presently occur in the northern Baie Verte Peninsula (Rowe 1972) and the mid-Holocene *Pinus* influx maximum (<300 grns·cm^{-2}·yr^{-1}) is too low to indicate that the genus was growing at the Compass Pond site as Davis (1976) says that 2000 grns·cm^{-2}·yr^{-1} indicates local arrival of the taxon. Thus the increase reflects the northward expansion of its range south of the Compass Pond site. In this context it is interesting to note that white pine, a typical species of a mixed forest, immigrated to northern Ontario and became an important component of the vegetation between ca. 7300 and 2600 BP (Liu & Lam 1985) and grew north of its present distribution in Québec ca. 5000 BP (Terasmae & Anderson 1970). That pines grow particularly well in unshaded sites and where natural fires are frequent (Watts 1979), may suggest that there were increased outbreaks of fire in northcentral Newfoundland.

Other evidence of a mid-Holocene hypsithermal is the occurrence of certain taxa which are either exclusive or nearly exclusive to this subzone, namely *Fraxinus*, *Taxus* and *Corylus*. Each of these taxa are represented by only one species in Newfoundland: *Fraxinus nigra*, *Taxus canadensis* and *Corylus cornuta*. In each case, the Newfoundland populations are at the northern limits of their range for eastern North America. The present distribution of *F. nigra* coincides with the warmest valleys of the island (Damman 1976), primarily south of Bonne Bay, and there is one record from the Springdale area (Ryan 1978). The presence of black ash, a very poor pollen producer, in the CPII diagrams suggests that either the populations southwest of Compass Pond were flowering profusely or that the small trees had migrated eastward (which implies that its occurrence at Springdale represents a relict distribution) or both. Given that its present distribution is controlled by the length and warmth of the vegetative season (Damman 1976), either of the above cases could best be explained by an ameliorating climate.

Similarly, the frequent appearance of *Taxus*, and to a lesser extent *Corylus*, during this period may represent population expansion or increased productivity. On the other hand, a more open forest canopy permitting wider dissemination of pollen grains may account for presence of these taxa during this stage. *Taxus* and small trees occurred on the Avalon ca. 8300 BP at a time when high 'tree' birch pollen was registered which Macpherson (1982b) interpreted as an open woodland with an understory or glades of shrubs. It is suggested here, rather, that shade-intolerant species such as *Corylus* (Davis 1976) were favoured by a mixed deciduous/coniferous canopy which allowed more light to infiltrate the forest and which also may have taken advantage of clearings resulting from recurrent fires.

A high sediment accumulation rate and the increasingly diverse nature of vascular aquatics suggests that lacustrine productivity may have been greater than earlier. Unfortunately, the value of these taxa as indicators of palaeolimnological conditions is limited because they comprise certain species that have different tolerances to water basicity and others that show no preference for particular chemical conditions.

(v) <u>*Picea-Betula-Alnus* Zone</u> (ca. 3200 BP to present)

An increase in spruce, at the expense of white birch, is apparent in the CPII profile after ca. 3200 BP, and is accompanied by an increase in alder. A resurgence in spruce has documented elsewhere in eastern North America and has been referred to as a stratigraphic marker for the end of the Holocene hypsithermal interval (Davis 1983). This 'boreal trend', expressed by an expansion of spruce populations, has generally been ascribed to a decrease in temperature or an increase in precipitation or both (Livingstone 1968; Mott 1975a; Anderson 1980; Davis 1983) and has been associated with an overall decrease in pollen influx.

In contrast, the CPII profile does not manifest an immediate decline in total pollen influx. In fact, the period between ca. 3000 and 2000 BP registers the highest pollen influx and sediment accumulation rates. The maximal total pollen influx values and sediment accumulation rates may reflect an increase in inputs to the lake or, alternatively, may result from sediment focussing, that is, differential deposition that results in greater net accumulation of sediment in the deeper parts of lake basins (Likens & Davis 1975). It is suggested that the total pollen concentration values during this period provide the answer. Studies have shown that sediment deposited in different parts of a lake tends to be homogeneous and well-mixed with similar pollen concentrations. Thus, the redeposition of sediment from one part of a lake to another would not alter the composition of the sediment although it would affect the rate of sediment accumulation and hence pollen influx. Therefore, since there is no appreciable change in total pollen concentration between 3000 and 2000 BP at CPII, it is postulated that the basin experienced changing patterns of sedimentation.

Davis <u>et al</u> (1984) have identified two processes that cause sediment focussing. The first mechanism is episodic and occurs when there is unusual weather with strong winds which create strong currents. The second is more regular and is associated with the annual turnover. As mentioned above, this boreal trend after ca. 3200 BP may have been due to cooler and moister conditions. If the increased precipitation was in the form of cyclonic storm activity, which could have lead to increased convective storm activity, during the summer months, then it would be likely that an increase in winds could provide the conditions necessary for intensified focussing.

It is interesting to note that, between ca. 3700 and 2000 BP, the influx curves and, to a lesser magnitude, the concentration curves of *Alnus*, *Picea* and *Abies* show several conspicuous peaks. In general, spruce peaks occur approximately 100 years after alder peaks. Fir peaks are more irregular, occurring either concurrently with spruce or 100 to 200 years later. The sample intervals during this period (ranging from 50 to 165 years) do not provide particularly fine resolution. Nevertheless, it is tempting to compare these results with those of Green (1981) who employed pollen and charcoal data to model post-fire responses of individual taxa. His study in Nova Scotia revealed that both spruce and fir pollen rose significantly about 50 years and 250 years, respectively, after a major fire. Green (1982) concludes that intense, widespread fires occurred in the early forests (11,000-6000 BP) in Nova Scotia when large tracts of land were dominated by inflammable conifers. It is plausible that major fire episodes occurred later near Compass Pond after black spruce became the dominant in the forest community. Also, periodic outbreaks of fire would have opened up the forest, promoting the spread of alder (Ritchie 1985).

Green (1981) adds, however, that other processes operate in a cyclic fashion to produce similar periodicities (e.g. budworm outbreaks increase fir mortality). Since charcoal peaks were not identified nor charcoal influx determined in this study, it is not possible to describe the frequency, intensity or extent of past fire events. However, the presence of charcoal in every sample in the CP-4 zone implies the continuing importance of fire in the development of the modern boreal forest. The above suggestion of increased moisture availability after 3200 BP does not negate this argument, for it is possible that an increase in the frequency of summer precipitation was accompanied by an increase in lightning storms.

A significant decrease in the rate of sediment accumulation, total pollen influx and, to a lesser degree, total pollen concentration is registered at CPII after ca. 2000 BP. A late-Holocene decline in pollen influx has been recorded at most sites in eastern Canada and has often been associated with a decrease in the vigour and density of the regional vegetation. In certain areas, changes in the representation of particular taxa have been interpreted as indicative of more open conditions. For example in northcentral Labrador, an overall increase in the tundra component was noted after ca. 3000 BP (Lamb 1985), and in the upland of northeastern Avalon, there occurred a relative increase in the shrub component after ca. 3200 BP (Macpherson 1982b). Yet in southeastern Labrador, Engstrom and Hansen (1985) state that changes in local sedimentation rather than terrestrial biomass changes are a more likely explanation for the observed concurrent decline in the influx values of all taxa after 3000 BP. However, their conclusions are contrary to those of Pennington (1979) who interprets concomitant decreases in influx and concentration as indicating real decreases in pollen inputs.

At Compass Pond, the percentage curves are, for the most part, uneventful although there is a small increase in fir, suggesting moister conditions, and a renewed representation of juniper which may indicate an expansion of peatlands, heaths and other exposed habitats (Ryan 1978). It is suggested that the significant decrease in pollen influx after 2000 BP is a combination of reduced pollen inputs to the lake reflecting continuing climatic deterioration, and decreased sediment focussing. That the accumulation of younger sediment in thinner layers often results in a Late-Holocene decrease in pollen influx has been documented for several lakes (e.g. Davis & Ford 1982; Davis et al. 1984).

8.4 Vegetation Reconstruction of Small Scrape Pond Site

(i) Betula-Cyperaceae subzone (ended ca. 10,400 BP)

Inspection of this subzone reveals that the fossil pollen assemblage has no modern counterpart. The dissimilarity is expected since modern tundra assemblages represent a comparatively static community whereas the SSP-1i assemblage represents an expanding pioneering community. Although it is difficult to find similarities with other fossil assemblages, this subzone bears some resemblance to, but is not contemporaneous with, the beginning of the CP-2i (*Betula*-Cyperaceae-shrubs) subzone.

Shortly after deglaciation, sedges colonized the area. Turbidity of the lake was sufficiently reduced to enable *Pediastrum* colonies to increase and flourish. As occurred at the Compass Pond site, the *Pediastrum* peak corresponds to a change in sediment lithology from clay to clay-gyytja. As expected, the subzone is floristically richer ($\bar{x}$ number of taxa identified = 21) than the earliest zones of CPII, for a more diverse flora was already nearby and available to colonize the newly exposed mineral substrate.

(ii) Betula-shrubs-Picea subzone (began ca. 10,400 BP)

Again the fossil pollen assemblage of this subzone lacks a modern analogue and represents a fortuitous assemblage of plants. This subzone resembles the *Betula-Picea* subzone of CPII in that there is a transition from a sedge-dominated community to one of dwarf shrubs and then a tall shrubs community interspersed with trees. The number of taxa identified is highest in this subzone ($\bar{x}$=23) as it is in the *Betula-Picea* subzone of CPII. The succession of dwarf shrubs at Small Scrape Pond differs slightly from that at Compass Pond. The sequence at the former was *Salix* and *Shepherdia canadensis* followed by Ericales and *Juniperus* whereas at the latter site the sequence was *Juniperus* followed by *Salix* then Ericales. Clubmosses and ferns were common during this vegetation stage.

Although the length of time covered by this subzone cannot be accurately determined, it is estimated that these vegetational developments occurred relatively sooner after deglaciation than they did at Compass Pond. This suggests that the development of the vegetation near Small Scrape Pond was temporally compressed which may be explained by the proximity of seed sources.

Populus pollen was not recorded at SSPII until after 10,400 BP and even then it never exceeded 1%. It is suggested that whereas propagules of the tree were able to spread rapidly and colonize areas that were deglaciated relatively early (e.g. Compass Pond), *Populus* was probably not able to compete well with the northward expanding populations of dwarf shrubs that rapidly colonized the Small Scrape basin.

The rapid increase in total pollen concentration is primarily due to the immigration or expansion of shrub birch (83% of the birch grains measured at level 281cm were <20μm). The sharp and concurrent rise in both the relative and absolute values of *Picea* indicate that spruce had arrived possibly as early as 9800 BP, but (as discussed in section 8.5) it was more likely 200 to 400 years later.

(iii) <u>*Betula-Picea-Alnus* zone</u>

This zone represents the arrival and expansion of all the major boreal forest species. The decrease in the shrub (excluding *Alnus*) and herb components point to the closing of the forest. The high concentrations of *Betula* and *Alnus* pollen during this zone are unlike those reported from the modern boreal forest (Davis & Webb 1975) and support evidence from the Compass Pond site that a white birch-black spruce forest dominated the regional landscape. In addition, the continuous presence of charcoal in samples above the birch peak of SSPII supports the hypothesis that recurrent fires were important in the regional forest history.

Although the frequency of *Abies* pollen is low, the increase to >4% at the base of this zone, and the continued occurrence of the grain thereafter, suggest that the tree arrived early in the zone. Because the sampling interval is coarser in SSPII than in CPII it is difficult to determine whether *Abies* preceded *Alnus* as appears to be the case at the Compass Pond site. That an *Abies-Alnus* peak is absent from SSPII may also be a result of the sampling interval.

Betula pollen percentages and concentrations continue to rise and culminate in a tremendous peak at 240cm. Grain measurements at this level reveal that only 7% were less than 20μm in diameter, although 90% may have derived from either shrub or tree sources. The increase probably represents the expansion of first 'shrub', and then 'tree', birch populations. Possibly this birch peak produced the peak in organic matter interpolated (from the SSPI profile) at ca. 7300 BP, which is over 1000 years earlier than at the Compass Pond site. It is suggested, however, that this date is too old and that the extremely high total pollen concentration is indicative of a rate of sediment accumulation that is considerably slower than the mean rate employed to determine the date of this event.

Discriminant analysis was not performed on *Picea* grains from this site. It is interesting to speculate, however, that the initial rise, the tapering off, and subsequent peak near the top of the SSPII profile may parallel the sequence of white and black spruce succession that occurred at Compass Pond.

Vascular aquatic vegetation developed in this zone and coincided, to a large degree, with the decline in *Pediastrum*. *Nuphar* was the first to become established following the deposition of organic mud on the lake floor. The increase in *Pinus* and *Taxus* pollen near the top of the profile (interpolated date of ca. 6000 BP from SSPI) may indicate the onset of the Holocene hypsithermal interval and supports similar evidence from CPII.

8.5 Migration of Dominant Forest Trees

An attempt is made here, using the seven dated pollen profiles from central and northcentral Newfoundland, (i.e. this study & Macpherson in prep.), to address questions pertaining to (a) the local arrival times of selected tree species at these sites, (b) the rate of species migration and seed dispersal distance, and (c) the directions of species movement. Although the density of the pollen sites is low, a preliminary isochrone map has been drawn to depict the diffusion of trees across part of the island (see Figure 8.2).

In order to determine the arrival time of a taxon near the site in question, it is necessary to distinguish between the locally-produced pollen and the far-travelled pollen components. The approach of Davis (1981) is to identify the time at which the curve for a pollen type exhibits a sharp increase. Others (e.g. Webb, Cushing & Wright 1983; Webb, Richard & Mott 1983), employing contemporary pollen-vegetation relationships, have determined the pollen frequency for each taxon which, they claim, indicates the taxon's presence. Bennett (1983, 1985) points out the difficulty in differentiating between the arrival of a species and its expansion at a site. He suggests that since a pollen curve rise frequently represents exponential population expansion of a species, it is safer to interpret the low, continuous pollen records as indicating local presence.

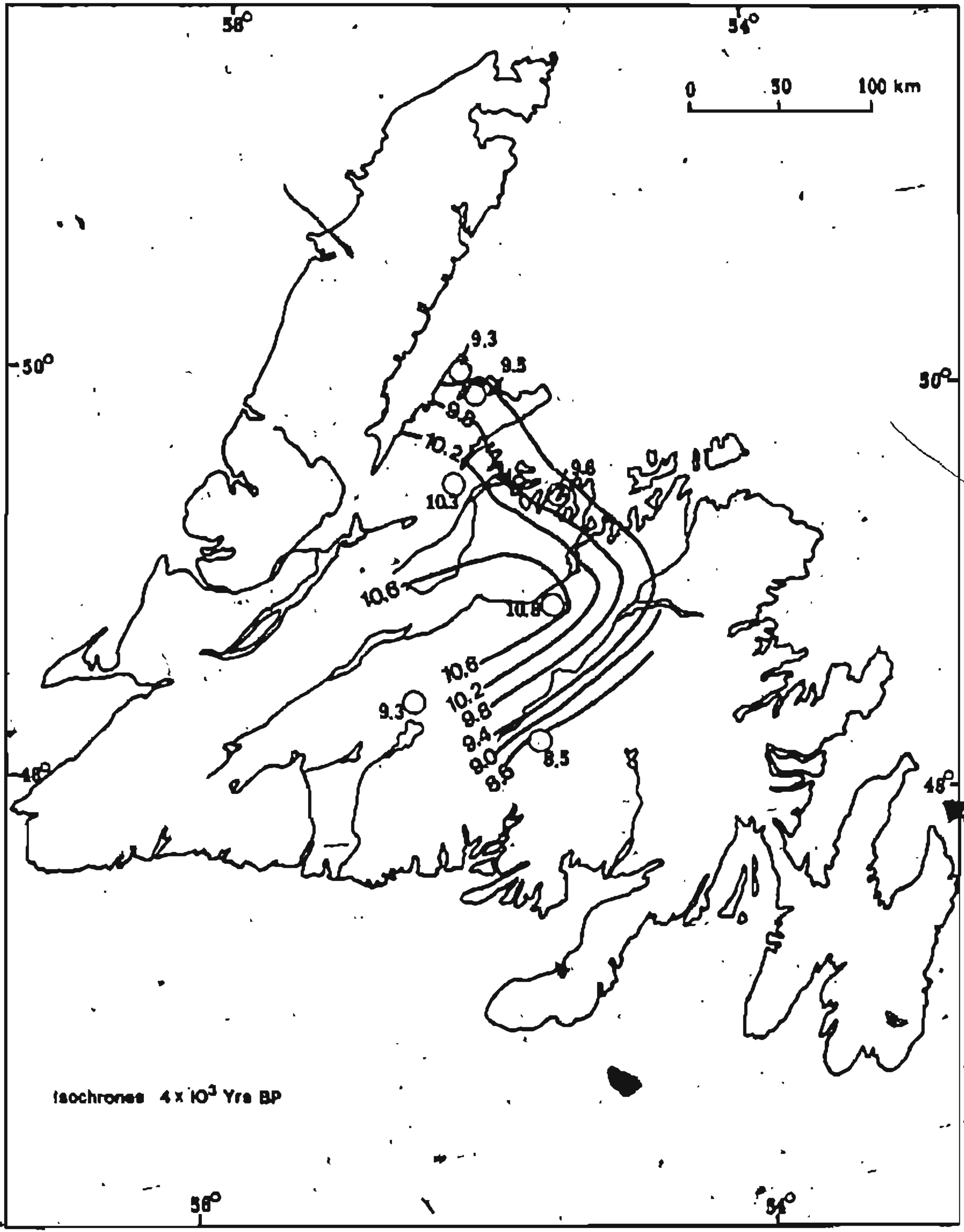

Figure 8.2: Isochrone map of Picea migration in central and northcentral Newfoundland.

It would appear that no one approach can be applied to all taxa given the differences in pollen production, dispersal and preservation. In this study, the concurrent rise in both the relative and absolute amounts of *Abies* pollen was interpreted as the local arrival time of the tree. Since *Picea* pollen is generally dispersed further from the parent tree than is the case with *Abies*, the low but fairly continuous representation of *Picea* before its rise was interpreted as far-travelled grains. On the preliminary time-distance diagram (Figure 8.3), the minimum age of glacial retreat and arrival times of spruce and fir are plotted for eight sites which roughly form a north-south transect covering 300km. Several of these events have been radiocarbon dated whereas the timing of others have been interpolated.

Picea

In the northern Baie Verte Peninsula the spruce rise is first recorded at SSP, yet interpolating the date of this event is not straightforward. The spruce rise occurs in the clay-gyttja whereas the fir rise occurs in the gyttja. Use of the mean sedimentation rate for SSPII would place the spruce rise at 9800 BP and the fir rise at 9500 BP (see Figure 8.4). However, the basal date of SSPI, within the clay-gyttja layer, is dated ca. 9500 BP. Therefore, a comparison of the basal dates and the stratigraphic position of the above vegetational events suggest that the clay-gyttja layer accumulated over a period of about 1200 years, placing the spruce rise closer to about 9500 BP. As spruce arrived approximately 200 years later at CP, its estimated migration rate was 75m yr^{-1}. This figure is comparable to the taxon's rate of diffusion in other parts of northcentral Newfoundland. Spruce advanced from the BF site at 10,800 BP arriving at the LT site, 65km to

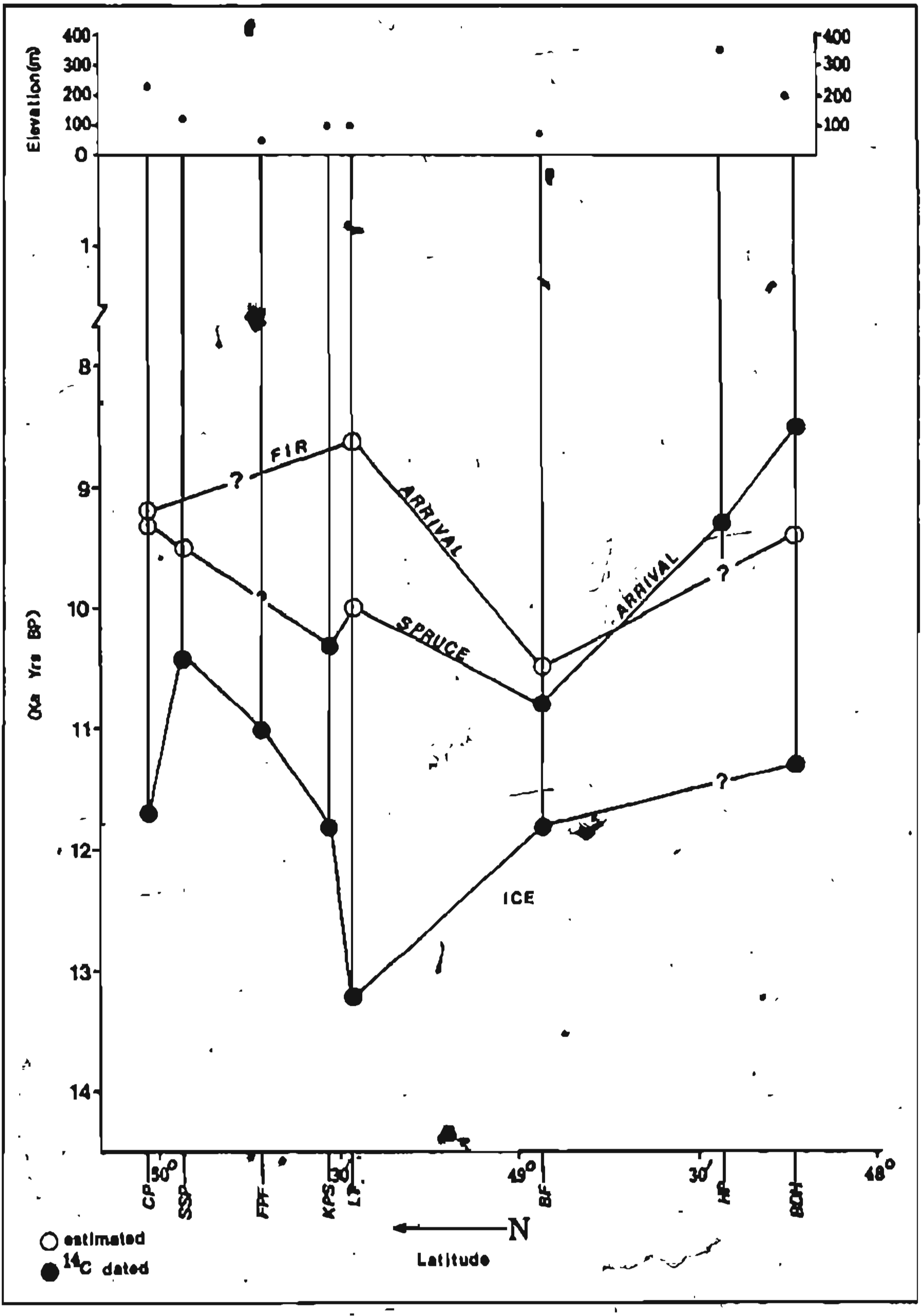

Figure 8.3: Chronosequence showing deglaciation and conifer migration in central and northcentral Newfoundland.

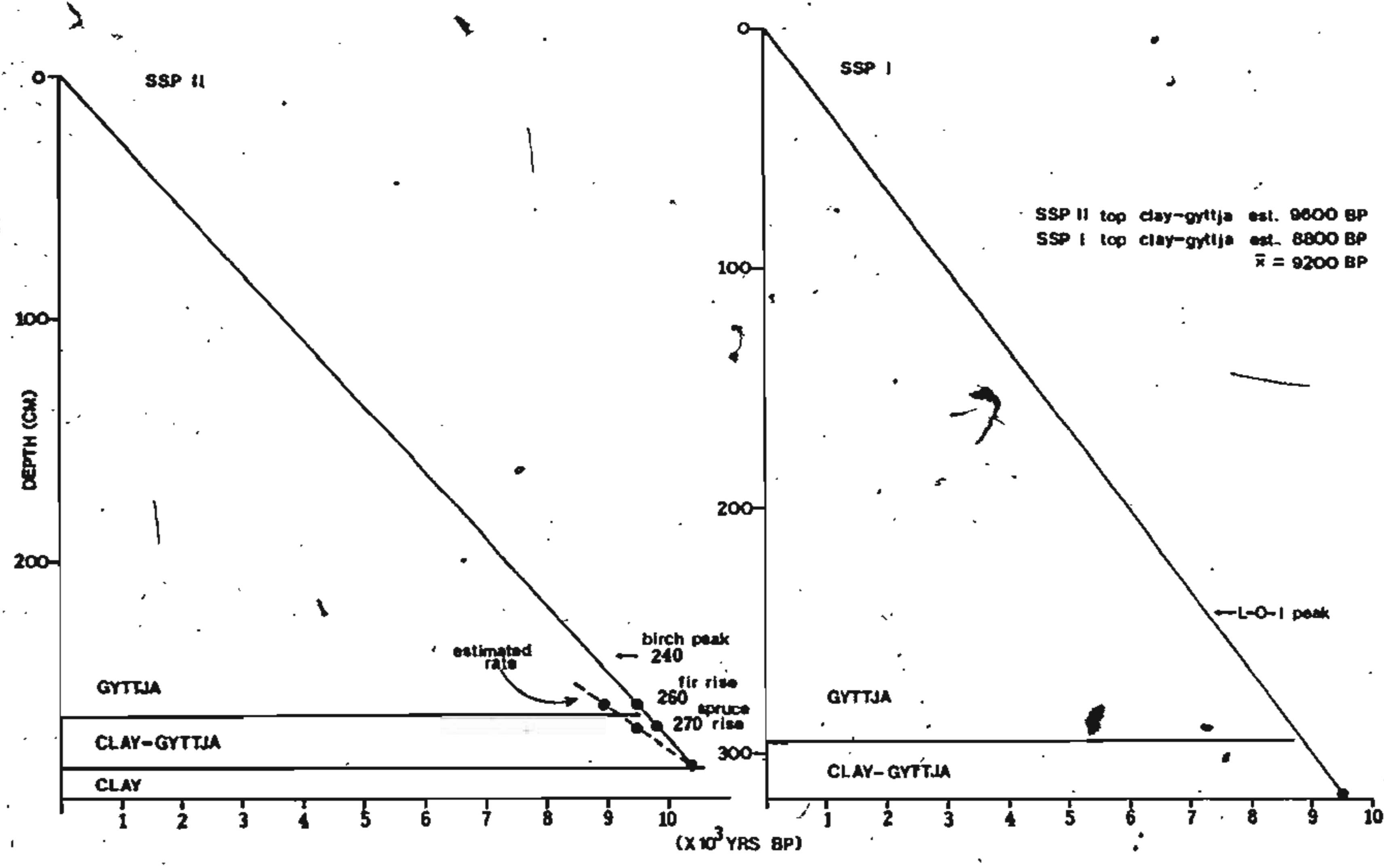

Figure 8.4: Small Scrape Pond I & II, Time-depth curves.

the north, at 9600 BP giving a migration rate of $54m \cdot yr^{-1}$, and spread southwards 75km to BDH at about 8500 BP, at a rate of $33m \cdot yr^{-1}$. From the KPS site (at 10,300 BP) it spread northward to CP (at ca. 9300 BP), a distance of about 65km, at a rate of $65m \cdot yr^{-1}$. Spruce did not arrive at the HP (Hidden Pond) site, in central Newfoundland, until 9300 BP (GSC-4186; Macpherson in prep.). However, this may be due to the elevation of the site which, at ca. 345m, is considerably higher than the other sites. These migration rates for spruce are considerably slower than those determined by either Davis (1981) for eastern North America (i.e. $250m \cdot yr^{-1}$) or Ritchie (1984) for northwest Canada (i.e. $1km \cdot yr^{-1}$).

If 10 years is taken as the minimum age at which spruce attains reproductive maturity in Newfoundland (Richardson & Hall 1973a), then the above rates of range extensions on the island imply an average seed dispersal distance from trees of approximately 650m. This is not far in excess of values cited by Ritchie (1984) for modern estimates of white spruce (i.e. 60-300m). Yet this may be an underestimate as white spruce does not usually seed until 20 years of age (Fowells 1965; Richardson & Hall 1973b). However, since the seeds of white spruce are shed annually in the fall (Ritchie 1984), the movement of seeds over frozen snow and ice surfaces may have been an important secondary dispersal mechanism.

The arrival of spruce at the SSP and BF sites occurred at least 1000 years after deglaciation and at the KPS site at least 1500 years after that area became ice-free. Its delayed arrival at CP (>2500 years following deglaciation) can best be explained by residual ice lingering in lowland areas of the Baie Verte

Peninsula. However, for the island in general, the spread of spruce in the wake of glacial retreat must have been rapid. Macpherson (in Blake 1983) points out that the main portion of the island's residual ice cap disappeared shortly before 11,400 BP, yet spruce had spread to the island and had arrived in central Newfoundland (BF site) by 10,800 BP. The dates of spruce immigration to sites in central and eastern Newfoundland provide a evidence of the route of its migration; the taxon appears first in central Newfoundland (at BF), later in northcentral Newfoundland, and later still on the Avalon Peninsula and southcentral Newfoundland (at BDH). The pattern strongly suggests that a main path of diffusion to central Newfoundland was along the Red Indian Lake - Exploits River valley. That spruce did not arrive until 9300 BP at the HP site in central Newfoundland does not negate this argument as the elevation of this site exceeds that of the BF site by 270m. Other valleys, including those of South Brook and Barney Brook, may have been important corridors for the northward advance of spruce to Notre Dame Bay and the Baie Verte Peninsula. The evidence thus points to the southwest portion of the island as the area from which spruce populations diffused.

Is it possible to estimate the arrival time of spruce in southwestern Newfoundland? Employing a conservative rate of 66m yr^{-1}, the average rate in northcentral Newfoundland, would provide a date greater than 15,000 BP, which predates the retreat of ice in the area (see Chapter 2), whereas using a rapid rate of 1km yr^{-1} (cf. Ritchie 1984) would result in a date greater than 11,000 BP. Both dates are far earlier than the earliest recorded arrival times of spruce in either Nova Scotia (ca. 10,800 BP (Hadden 1975)) or Prince Edward Island (ca. 10,000 BP (Anderson 1980)).

An exceptionally early arrival of spruce in southwestern Newfoundland would imply that either the continental shelf, rather than the Maritimes, acted as a corridor for northward migrating trees, or that spruce survived the last glaciation in a coastal refugium, possibly the southern Burin Peninsula, or close by on the exposed continental shelf. Both black and white spruce can persist as prostrate krummholz, reproducing vegetatively, beyond the limit of tree growth ([illegible]ette & Gagnon 1979; Ritchie 1984). It appears unlikely that spruce could have arrived much later than 11,000 BP in southwestern Newfoundland for that would have involved exceptionally great seed dispersal distances. Even a migration rate of $1 km\,yr^{-1}$ implies a distance of spread per generation of 10km which far exceeds modern estimates. However, since spruce most likely migrated northward along the Exploits River valley, its spread could have been rapidly accelerated by the river transport of seeds and even branches.

Autecological studies of black spruce indicate that estimated minimum average monthly temperatures of 9.5°, 13° and 10° for June, July and August, respectively, are necessary for successful reproduction (Black & Bliss 1980). Thus the early presence of spruce in central Newfoundland and the implied northward extension of its range in the interval ca. 11,000 to 9300 BP, during which time the late-glacial climatic reversal was recorded at LT and the southern Burin, serves to contrast the climates of the interior with coastal areas.

Clearly further sites, particularly in southwestern Newfoundland, need to be investigated. At present it is only possible to hypothesize that spruce migrated to the island during the initial period of warming, before ca. 11,300 BP. In addition, it should be emphasized that this spread of spruce was not at high population densities but rather represented scattered populations. Studies of Holocene plant migrations have shown that the distributions of many species were in disequilibrium with the climate and that numerous other mechanisms, including propagule availability, longevity of source plants, seedling competitive ability and availability of suitable migrational pathways played important roles in species migration (cf. Davis 1981; Bennett 1985).

Abies balsamea

The local arrival times of balsam fir are all based on interpolated dates. With the exception of the BDH site on the central plateau, fir invaded the sites after spruce had arrived, occurring first at the BF site at 10,500 BP, reaching the BDH site at possibly 9400 BP, the CP site by ca. 9200 BP and did not arrive at the LT site until ca. 8600 BP nor the Avalon Peninsula until 8300 BP. Its time of arrival at SSP is problematic. Although the pollen sequence indicates that spruce preceded fir, probably at about 9500 BP, it cannot be established whether fir arrived first at this site or at the CP site. As yet there is no date for the arrival of fir at KPS. The pattern of fir arrival times, though less complete than that of spruce, suggests a similar migrational pathway from southwestern Newfoundland. Presumably factors other than climate (e.g. competition, site availabilty, soil conditions) were important controls in its delayed arrival at certain areas (e.g. Notre Dame Bay). In eastern North America, the average Holocene migration rate of fir is estimated at 200m yr^{-1} (Davis 1983). Although the sites with dates are too scattered to determine the migration rate of fir within Newfoundland, it is suggested that the rate of spread of this tree across the island was slow compared to its spread in other parts of eastern North America.

Betula

Attempts to establish the arrival time of *Betula papyrifera* are complicated primarily because the pollen curves for *Betula* actually record the invasion of both shrub and tree birch. The task of distinguishing between locally produced pollen and far-travelled pollen is further complicated since most shrub and tree species of birch are prolific pollen producers and are often overrepresented (Birks & Birks 1980; Ritchie 1984). Only the immigration and expansion of *Betula* in northern Baie Verte Peninsula are discussed here.

In Figure 8.5, the abundance of *Betula* pollen (both percentage and influx) is plotted together with mean pollen grain size and range against depth for CPII. Does the initial rise in the relative and absolute amounts of *Betula* pollen represent the migrating front of shrub birch within the region, or indicate that birch had already arrived in the vicinity and that its population was expanding? Studies from Baffin Island show that where isolated groups of dwarf birch occur, birch pollen percentages exceed 5% (Andrews et al. 1980; Short et al. 1985). It would appear safe to say that in CPII the increases in percentage (from 7 to 21%) and concentration (from 230 to 2400 grns cm^{-3}) that occurred ca. 11,600 BP reflect the local arrival, if only at a very low population density, of shrub birch.

The exponential increase in *Betula* pollen influx after this date is interpreted as a change from a low to high population density culminating at ca. 9500 BP. The decline in percentage and, to a lesser degree, concentration values after ca. 8600 BP reflects a real decrease in the population of shrub birch, probably resulting from competition with other shrubs and trees. At ca. 7900 BP there is an increase in the mean size of birch pollen (Figures 8.5 & 8.6). It is suggested

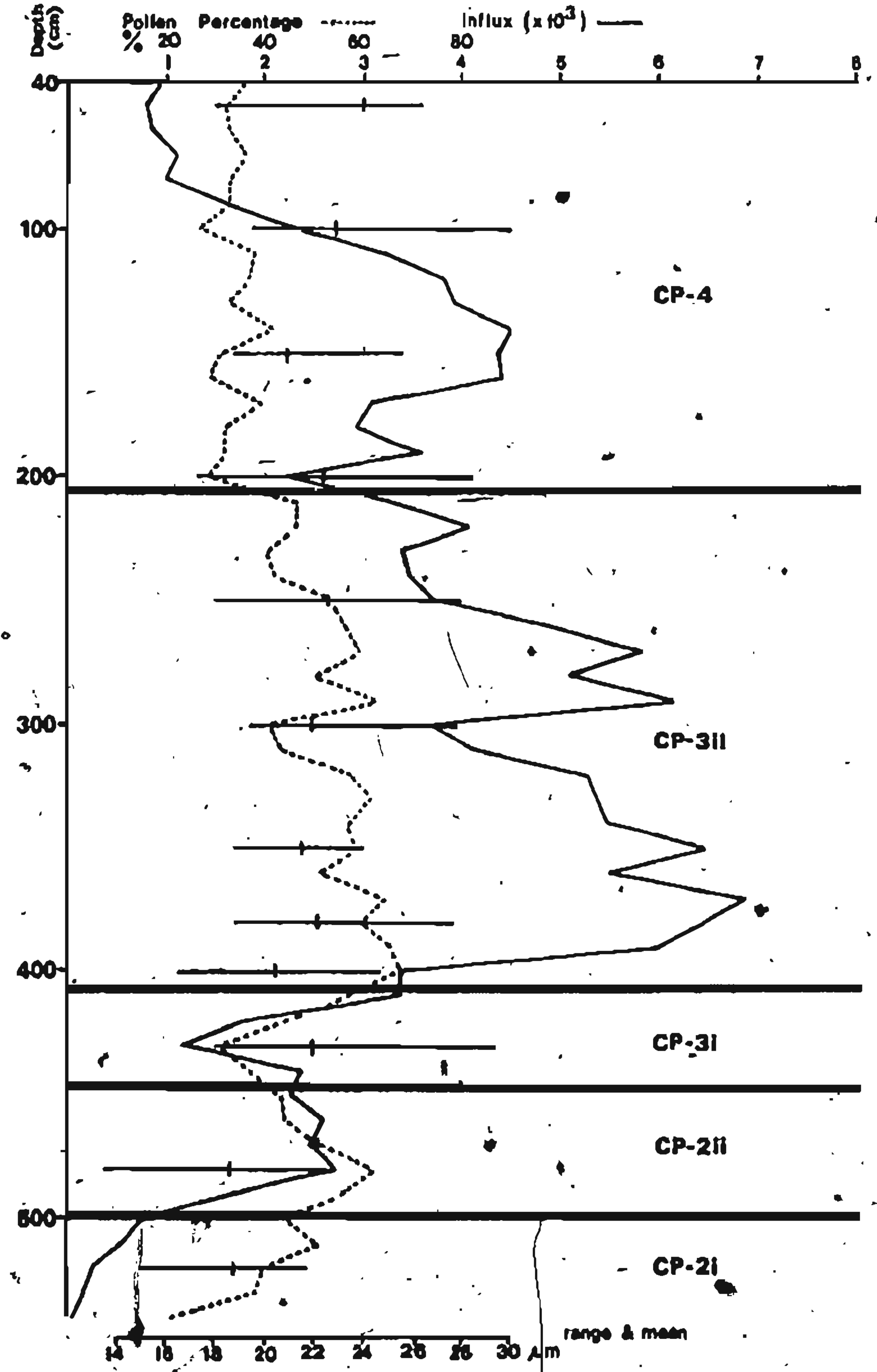

Figure 8.5: Compass Pond II, Betula grain size and abundance vs. depth.

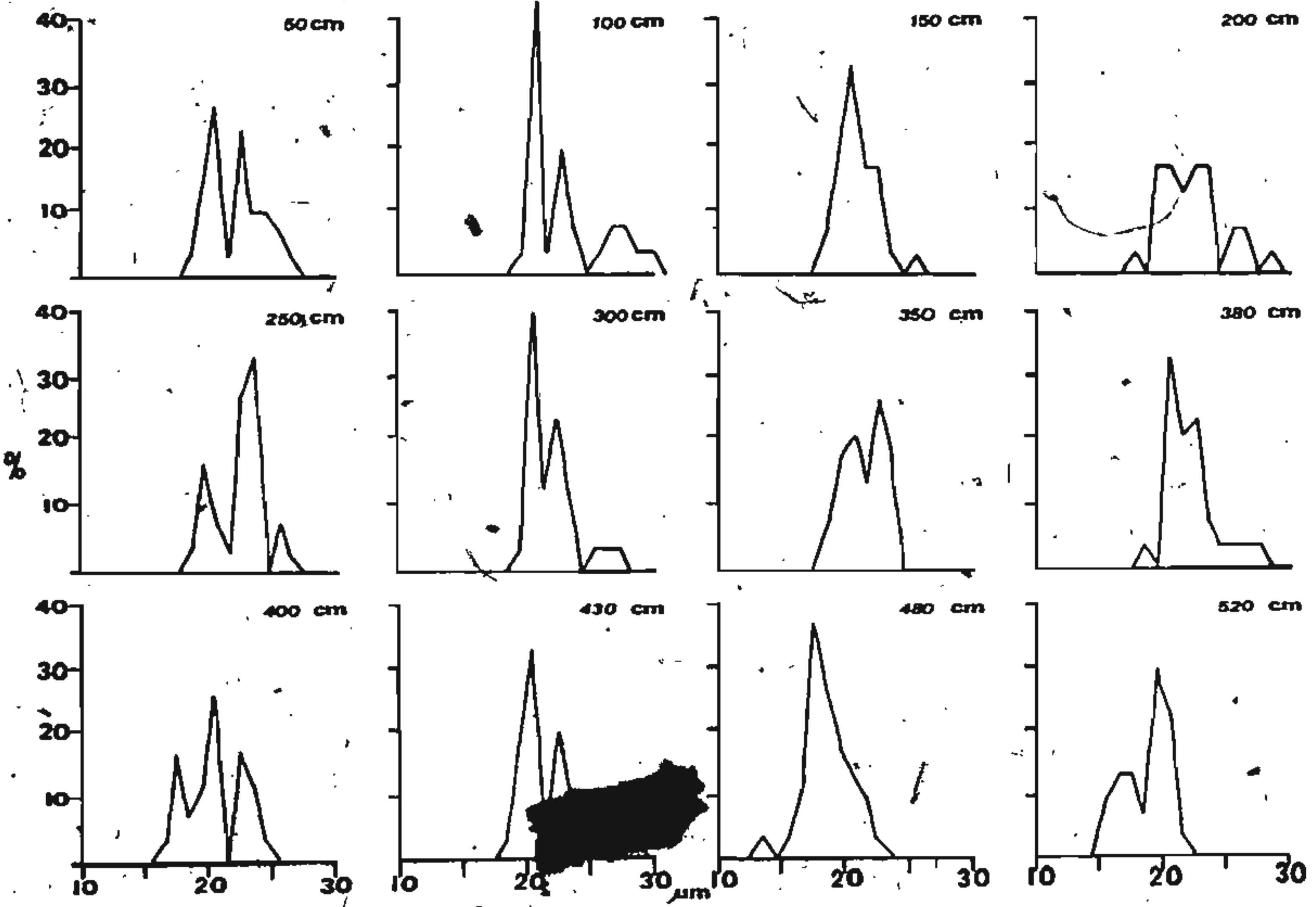

Figure 8.6: Compass Pond II, Betula grain size (polar view) distribution.

that at this time the contribution of shrub birch within the vicinity was low and unable to mask the input of either far-travelled or local tree birch grains. It is impossible to determine the arrival date of tree birch with precision but the data (see Figures 8.5 & 8.6) indicate that the tree most probably grew in the area before 6000 BP.

The SSPII diagram does not show the same pattern of birch pollen abundance as does that of CPII (see Figure 8.7). Thus, it is not improbable that a few individuals of shrub birch grew near this site immediately following deglaciation, ca. 10,500 BP. Although the mean grain size indicates that shrub birch was the main contributor to the *Betula* pollen curve after 10,400 BP, the range in grain size is significantly broader than that registered at CPII at about the same time. This suggests that whereas shrub birch was established at the CP site for a comparatively long period, at the SSP site either both 'types' of birch immigrated concurrently or tree birch followed shortly after the arrival of shrub birch. If it is accepted that the percentage and concentration peaks at 240cm in SSPII represent a peak in tree birch populations (See Figures 8.7 & 8.8), then either of the above scenarios would explain the absence of a decline and subsequent rise in birch pollen abundance, as is evident in CPII.

8.6 Summary of Vegetational Development and Implications for Holocene Climatic Change

Climatic change was the ultimate mechanism responsible for the post-glacial expansion of species ranges and long-term vegetation dynamics. However, the rate of species spread and the composition of communities that formed over time resulted from a complex interplay of both environmental and biological factors. The interaction of these controlling factors upon the vegetational development in

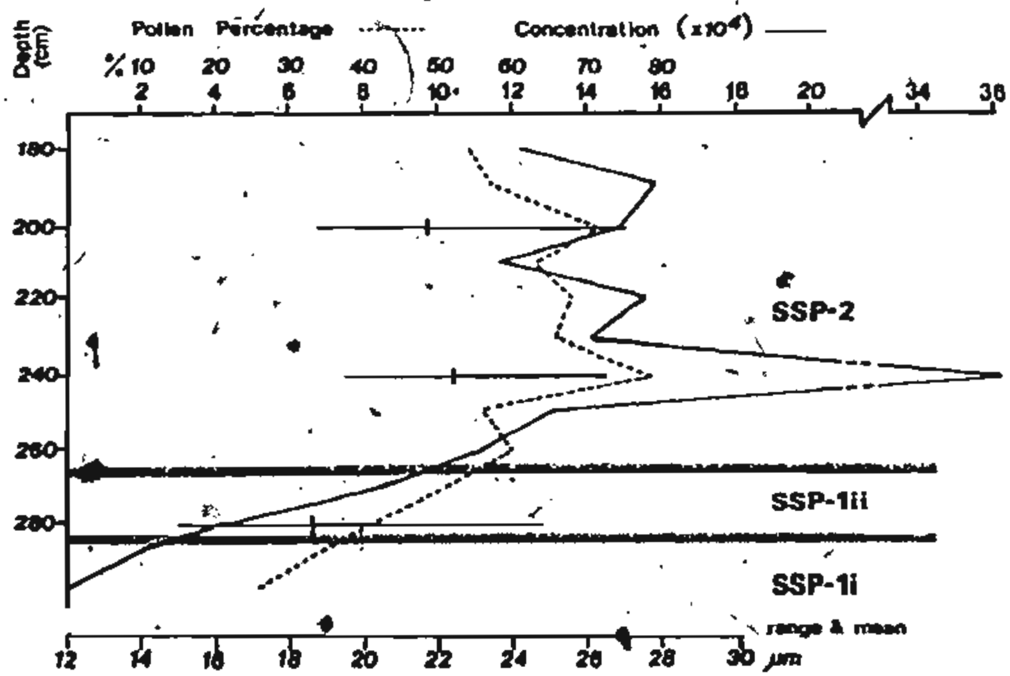

Figure 8.7: Small Scrape Pond II, Betula grain size and abundance vs. depth.

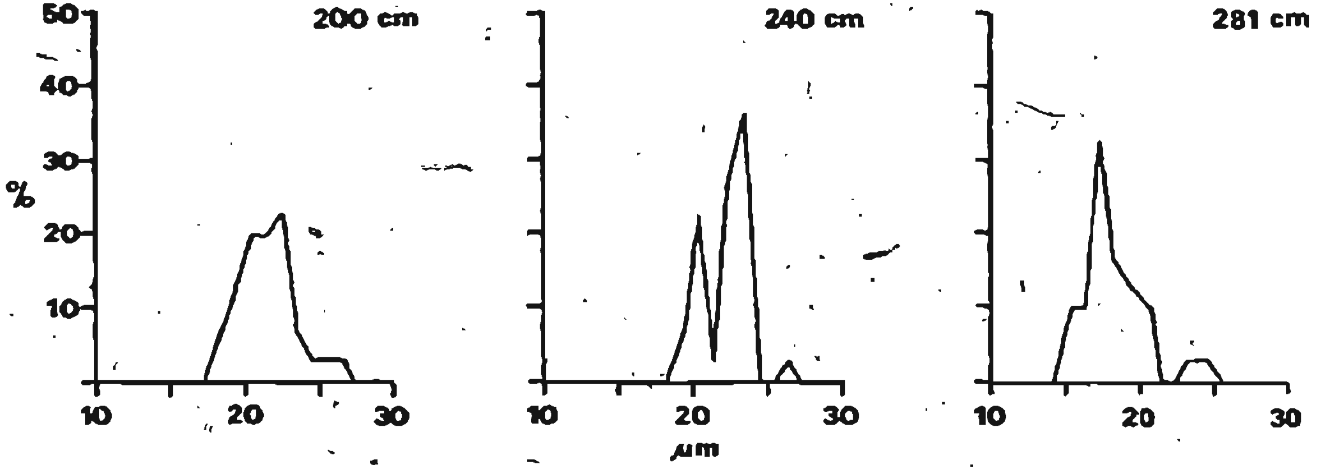

Figure 8.8: Small Scrape Pond II, *Betula* grain size (polar view) distribution.

the northern Baie Verte Peninsula is discussed below. Figure 8.9 has been drawn to facilitate a regional comparison of inferred climatic trends.

Deglaciation of the highland area of northern Baie Verte Peninsula occurred before 11,800 BP. Compositionally the herb-dominated initial communities were akin to the low-arctic steppe of western Greenland, but the sparseness of the vegetation probably resembled a high-arctic tundra. Wright (1984) claims that basal herb pollen zones may not represent tundra conditions, in the climatic sense, but rather a pioneering phase reflecting the proximity of seed sources. It is proposed that both reconstructions are true for the CP site, that is, a pioneering community colonized the landscape when the local climate was cold and dry. Whereas ice wastage on the Peninsula probably resulted from precipitation starvation and an ameliorating climate, the residual ice may well have affected the local climate (cf. King 1985a).

After ca. 11,800 BP the development of a sedge and dwarf-shrub tundra was completed by long-distance seed dispersal as lowland areas were still glaciated. It is difficult to discern whether an abrupt or gradual warming occurred after ca. 11,800 BP. Possibly the long shrub-tundra period (over 2000 years) reflected a climate too cold for tree growth. Alternatively, the length of the shrub-tundra phase may reflect the availability of both propagules and suitable migrational corridors for tree colonization. The latter hypothesis is favoured given that *Populus*, a taxon with a highly efficient dispersal mechanism, appears to have grown in the area shortly after deglaciation of highland areas.

There is no lithologic or pollen evidence of a late-glacial climatic oscillation. However, reversals in both loss-on-ignition and *Pediastrum* curves of the CP sites,

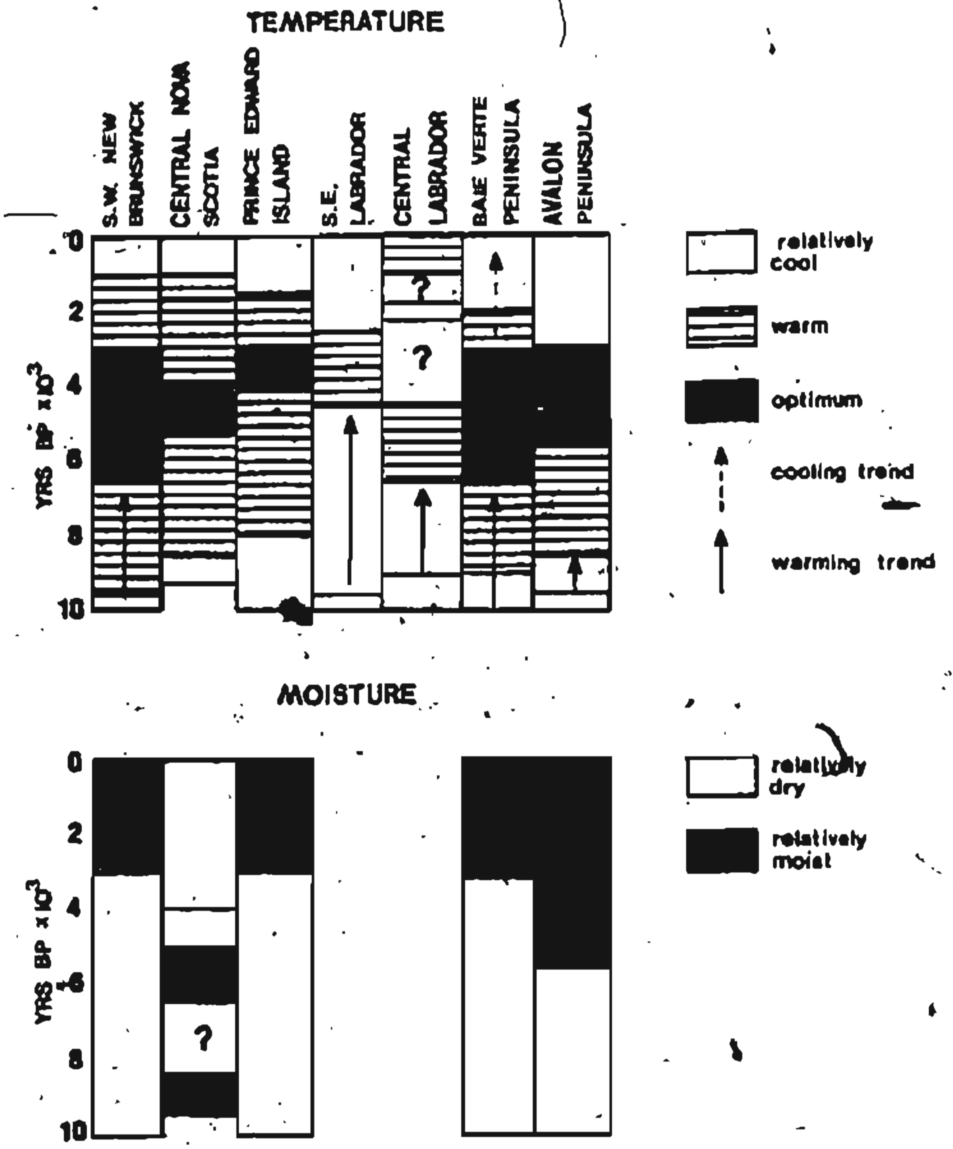

Figure 8.9: Inferred regional Holocene climatic trends.

occurring approximately between 11,300 and 10,500 BP, may be indicative of cooler conditions. It is noteworthy that loss-on-ignition data from the LT site show a decline during the inferred climatic reversal. In addition, this decrease in organic matter coincides with the initial decrease in *Pediastrum*.

Before ca. 10,400 BP, the SSP site had become ice-free and sedges were abundant. Populations of low shrubs, including *Salix*, *Shepherdia canadensis*, Ericales and *Juniperus* rapidly expanded into newly deglaciated areas. Low shrub communities were subsequently replaced by shrub birch and *Myrica* which became important components of the vegetation after ca. 10,000 BP.

Arboreal immigration and invasion of the shrub-tundra commenced ca. 9,500 BP with *Picea*, primarily white spruce, followed by *Abies balsamea*. Alder shrub populations also expanded. An unstable period continued until ca. 6700 BP. Presumably the landscape was covered with an open spruce-fir forest with an understory or glades of alder shrubs. Populations of white spruce and fir peaked and declined in turn in favour of black spruce and white birch.

It would be difficult to argue that the successive arrivals of dominant forest trees were a direct result of climatic change. For although an increase in fir at the expense of white spruce may have resulted from continued climatic warming, a subsequent increase in black spruce and decline in fir would imply a cooling trend (Engstrom & Hansen 1985). There is no corroborating evidence for this. In fact, pollen records from eastern Canadian sites indicate continued warming. Also, the rapidity of the successive peaks of white spruce and fir do not imply climatic control. It is proposed that the distribution of tree species was in disequilibrium with the climate and that the timing of these events can best be explained by other environmental and biotic factors, particularly changing edaphic conditions and species competition. Fires also probably contributed to the competitive replacement of white spruce and fir by black spruce (cf. Engstrom & Hansen 1985). This claim is partially supported by Damman who, in describing the vegetation changes from south to north in the Central Newfoundland Ecoregion, says that there is "a replacement of *Abies balsamea* by *Picea mariana* as the dominant tree in the landscape...due to the more important role of fire in the northern parts" (1983, p.175).

The closing of the forest canopy occurred after ca. 6700 BP. A white birch-black spruce forest developed and was maintained for approximately 3500 years. It is suggested that conditions were dry, that summers were warmer and the growing season longer than at present. Interestingly, Lamb (1980) infers a rapid amelioration of the climate in southeastern Labrador after 6500 BP. There was an expansion of *Pinus* populations south of the Peninsula between ca. 6000 and 4000 BP. This climatic reconstruction is closer to parts of the Maritime Provinces, in that the inferred period of optimum warmth corresponded to dry conditions (e.g. Mott 1975a; Anderson 1980), than for the Avalon Peninsula where it has been suggested that the period of maximum warmth (ca. 5500-3000 BP) was accompanied by an increase in precipitation (Macpherson 1982b, 1985). Also, the warmer than present period inferred for the northern Baie Verte Peninsula corresponds closely with the oceanic optimum which lagged behind the atmospheric optimum by about 1500 years, ending ca. 3000 BP (Andrews 1972).

A resurgence of spruce and decrease of white birch after 3200 BP indicates the onset of a cooling trend. Furthermore, it is suggested that more frequent summer precipitation or lower evapotranspiration accounted for the renewed importance of balsam fir after 2000 BP (cf. King 1985a). This reconstruction of a deteriorating climate corroborates evidence from the Avalon Peninsula and southern Labrador where a cooling trend occurred after 3200 BP and 2500 BP, respectively. In this regard, it is interesting to note that Davis (1984) concludes that the most pronounced episode of paludification in Newfoundland occurred ca. 3000-2000 BP, resulting in decreased forest cover. A decrease in abundance and sediment accumulation rate and the recurrence of certain shrub taxa after 2000 BP implies more open conditions. However, it is suggested that the considerable decrease in pollen influx at that time was enhanced by decreased sediment focussing.

8.7 Further Research

At least as many questions have evolved from this study as it has attempted to answer. The following are some areas to which future palynological investigations might be directed.

1. No routine counts of charcoal occurrence have been made on cores from Newfoundland sites. Such an analysis would likely prove invaluable in determining the importance of fires on Holocene forest development.

2. Analysis of the central Newfoundland cores, near Red Indian Lake, should provide more information on the arrival times of dominant forest trees. In particular, a study of the time of local arrival of spruce and fir in southwestern Newfoundland would yield important data relating to the migration rates and corridors of movement of these species across the island. Such a study might also throw light on to the location of glacial refugia of these species.

3. If discriminant analysis were performed on spruce grains as a routine procedure in palynological work, it might be possible to determine the sequence of white and black spruce immigration and their migrational pathways.

4. The lack of modern analogues presents a problem for the palaeoclimatic interpretation of pollen data. It is suggested that further studies need to be undertaken in Newfoundland to provide a clearer picture of modern pollen-vegetation relationships.

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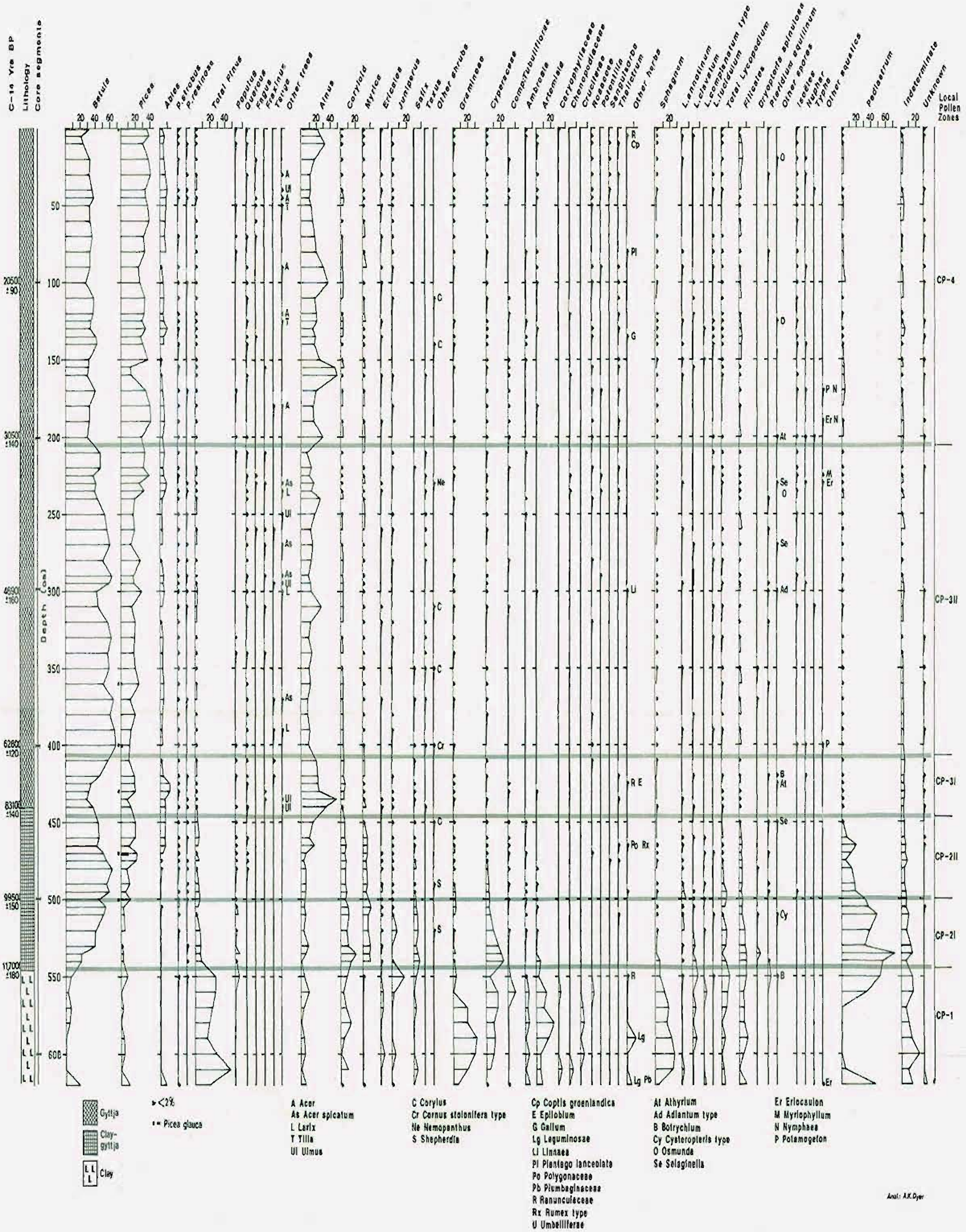


Figure A.1: Compass Pond II (50°02'03"N 56°11'47"W ca. 236m.a.s.l.)
Pollen Percentage Diagram

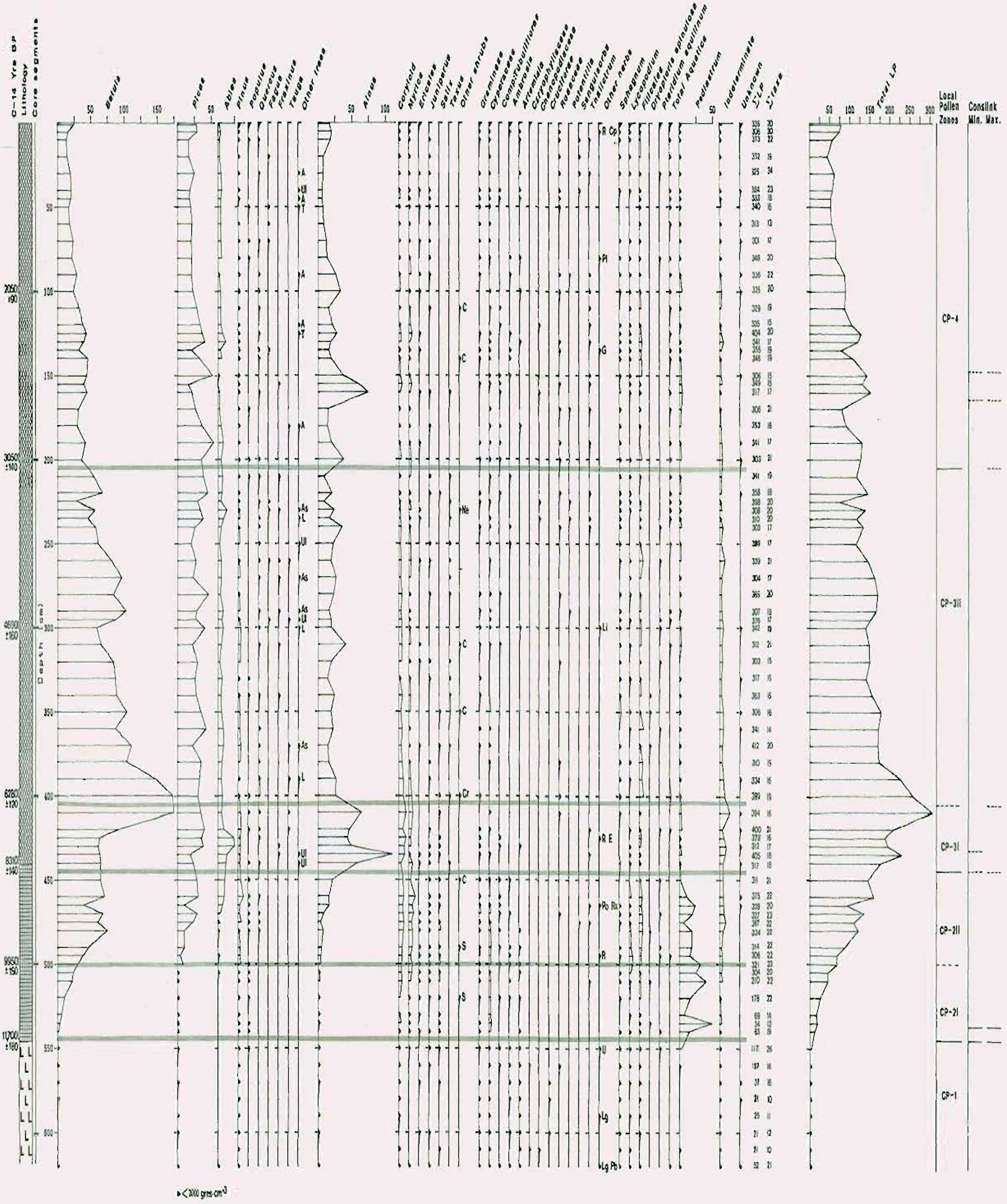


Figure A.2: Compass Pond II (50°02'03"N 56°11'47"W ca. 236m.a.s.l.)

Pollen Concentration Diagram ($x10^3$ grns·cm^{-3})

Anal: A.K.Dyer

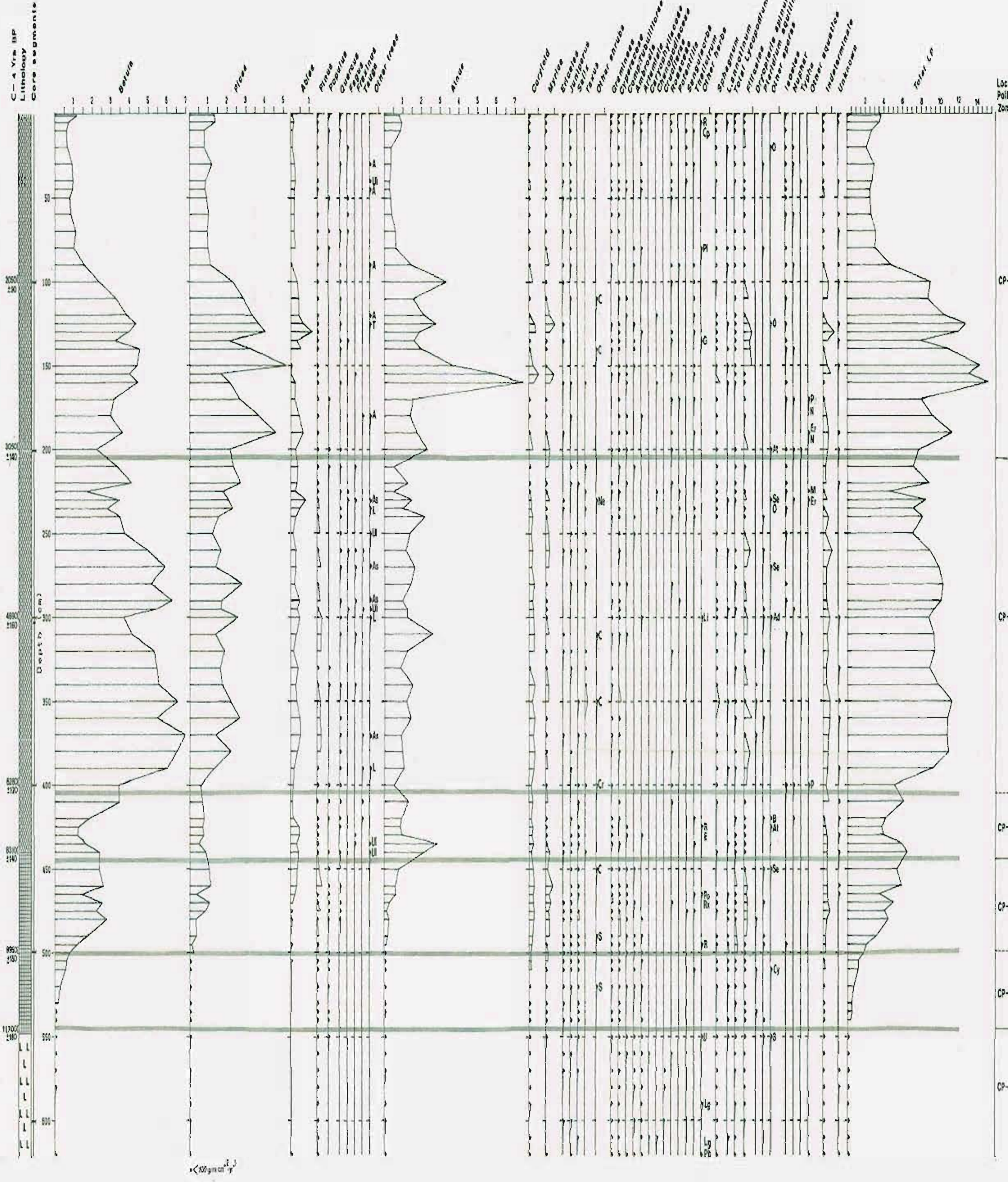
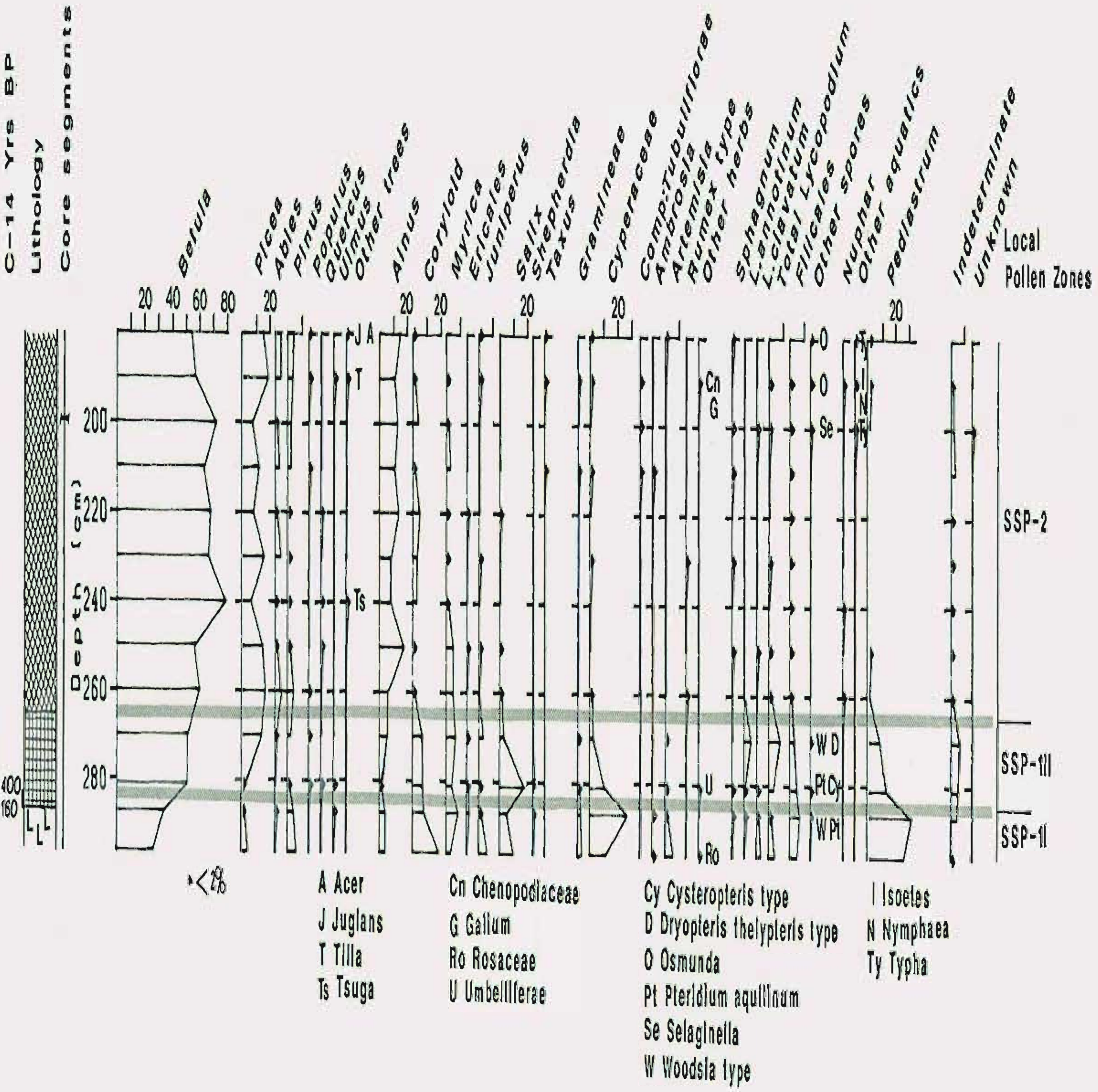
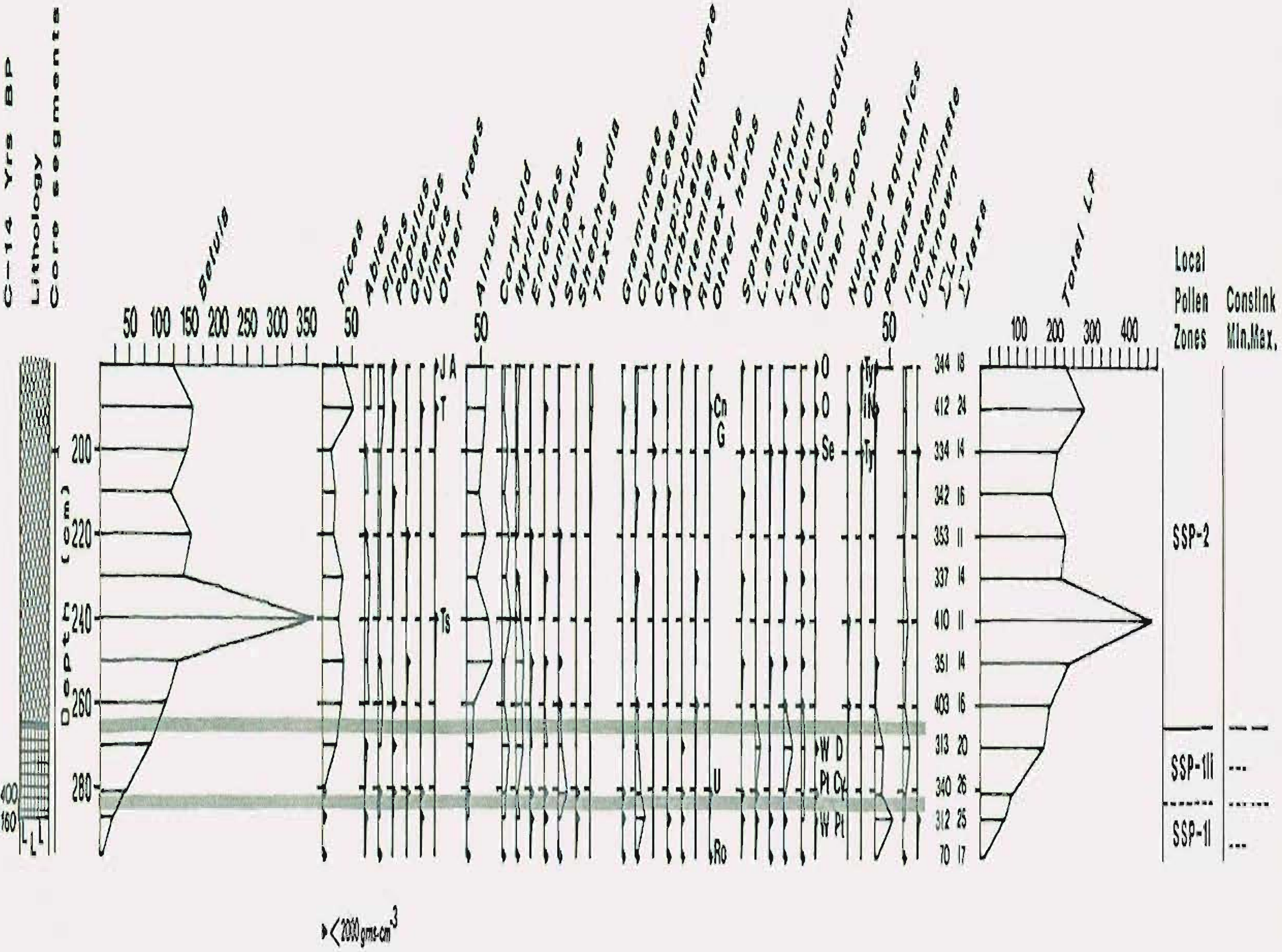


Figure A.3: Compass Pond II (50°02'03"N 56°11'47"W ca. 236m.a.s.l.)
Pollen Influx Diagram ($\times 10^3$ grns·cm^{-2}·yr^{-1})



Anal: A.K.Dyer

Figure A.4: Small Scrape Pond II (49°56'59"N 56°05'17"W, ca. 122m.a.s.l.)
Pollen Percentage Diagram



Anal.: A.K.Dyer

Figure A.5: Small Scrape Pond II (49°56'59"N 56°05'17"W, ca. 122m.a.s.l.)

Pollen Concentration Diagram (x10^{3}grns·cm^{-3})