PALEOCEANOGRAPHIC RECONSTRUCTION OF THE INNER SCOTIAN SHELF DURING THE YOUNGER DRYAS: A BENTHIC FORAMINIFERAL STUDY

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ABSTRACT

Examination of foraminifera from piston core 91-018-53P permits a paleoceanographic reconstruction of the Inner Scotian Shelf, with emphasis on the Younger Dryas cold period. A minimum in the total number of foraminifera and foraminiferal species, and the dominance of *Elphidium excavatum* forma *clavatum* characterize the Younger Dryas interval. These results suggest a warm ice margin during the Younger Dryas interval, confirming the presence of a proximal glacial ice mass. The representation of the Younger Dryas in core 91-018-53P, dated at 10,377 to 9,840 ^{14}C ybp, is the first well documented and dated marine occurrence on the east coast of North America. Additionally, the core contains the first known marine occurrence of the recently discovered Killarney Oscillation, which represents a minor cold period that pre-dates the Younger Dryas. The late Holocene (represented by *Bulimina marginata, Cassidulina laevigata,* and *Fursenkoina fusiformis* as well as peak occurrences of ostracods and planktonic foraminifera) and the early post-glacial (represented by *Elphidium excavatum* forma *clavatum, Cassidulina reniforme,* and *Islandiella teretis*) are similar to those previously reported from this area.

Key Words: foraminifera, paleoceanographic reconstruction, Younger Dryas, meltwater

I dedicate this thesis to my parents, not only for their support during the past year, but for their unending support during my four years at Dalhousie University.

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CHAPTER 1 INTRODUCTION

1.1 Introduction

Foraminifera are unicellular organisms of the Phylum Protozoa, subdivided into classes according to the type of locomotor apparati present. Foraminifera are non-flagellate, but instead have projections of protoplasm, termed pseudopodia, which places them into the Class Sarcodina that includes the simplest of the Protozoa with respect to their cellular organization and specialization (Boersma 1978). The presence of a mineralized shell (or test) generally composed of porcellaneous calcite, hyaline (clear) calcite, or agglutinated foreign particles with varying types of cement distinguishes foraminifera from other Sarcodina (Whitten and Brooks 1972).

Benthic foraminifera behave as biostratigraphic, paleoecologic, and paleoceanographic indicators. Their minute size, abundant occurrence, and wide geographic distribution in sediments of almost any age or marine environment enhances the practical value of these marine microfossils. The occurrence of assemblages of benthic foraminifera are environmentally-dependent and, because foraminifera are abundant and quantitative studies require only a small sediment sample, foraminiferal research has major applications in various fields of historical geology (Berggren 1978).

The field of Quaternary paleoceanography generally focuses on global-scale problems. Most of these large-scale studies deal with observations derived from deep-sea sediments providing long continuous records of marine microfossils, with minimal influx of clastic sediment. However, deep-sea records also present two problems : (1) slow sedimentation rates greatly condense stratigraphic records; and (2) they do not reflect immediate ice margin changes or deglaciation events (Scott *et al.* 1984).

Continental shelf records permit examination of the direct effects of glaciation on marine environments. This type of examination is often difficult because many continental shelves are currently areas of erosion, and obtaining an undisturbed continuous glacial-interglacial sedimentary record is virtually impossible (Scott *et al.* 1984). The seabed topography of the continental shelf of eastern Canada, with shallow bank areas and over-deepened intrabank basins commonly more than 300 m deep (Scott and Medioli 1988), is distinct from almost any other continental shelf, including that just to the south off New England (Scott *et al.* 1989b). These basins are sediment traps characterized by high sedimentation rates (up to 10 cm / century), resulting in expanded sedimentary and micropaleontological records (Scott *et al.* 1984).

1.2 Purpose and Scope

The purpose of this project is to study foraminiferal assemblages from piston core 91-018-53P (core location shown in Figure 1.1) to determine the bottom-water paleoceanography. The expanded sedimentary record allows for a detailed foraminiferal analysis with emphasis on the Younger Dryas cold period from 11,000 to 10,000 ybp (years before present).

The scope of this research includes foraminiferal, sedimentological, and seismic constraints, as well as radiometric carbon-14 ages of the sedimentary sequence represented in the 8 m piston core.

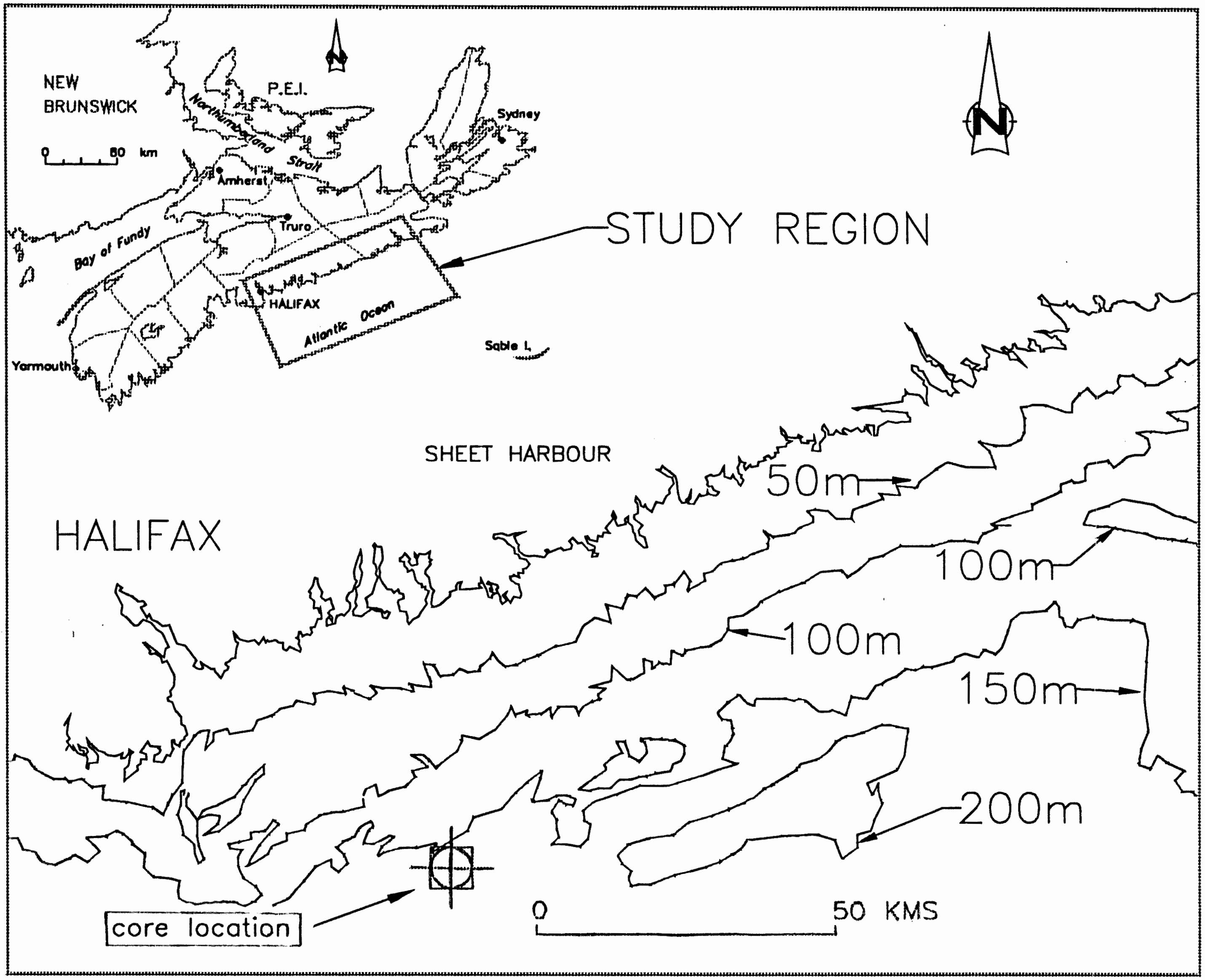


Figure 1.1 Approximate location of core 91-018-53P.

1.3 Physical Setting

The Scotian Shelf consists of a well-defined, rough inner shelf which is an offshore continuation of the land area (King and Fader 1986). The three main sediment types include muds, sands, and gravels. The offshore core site, approximately 46 km south of Jeddore Harbour in 168 m water depth (Fig. 1.1), is in a trough 2.5 km landward of the Halifax Moraine (the furthest extent of the most recent glaciation).

A southwesterly flow parallel to the coast, known as the Nova Scotia Current, characterizes the mean circulation over the Inner Scotian Shelf (Forbes *et al* 1991). Axially located 40 to 50 km offshore (Forbes *et al* 1991), it is weakly stratified with colder water of lower salinity at the surface near the coast and warmer, but higher salinity, water below and seaward (Forbes *et al.* 1991). The current, shown in Figure 1.2, dominated by the colder Cabot Strait (Gulf of St. Lawrence) waters, contains considerable additions of water from the Inner Labrador Current. Established from Cabot Strait surface water, the Inner Scotian Shelf bottom water has a salinity of 31.8‰ and a temperature range of 3-6°C (Williamson 1983).

1.4 Previous Investigations

The Geological Survey of Canada mapped the surficial sediments near the core site, complete with distribution of shelf lithologies and sediment facies (King 1970; and King and Fader 1986). Following this, Piper *et al.* (1986) completed an investigation of the marine geology off the South Shore, Nova Scotia.

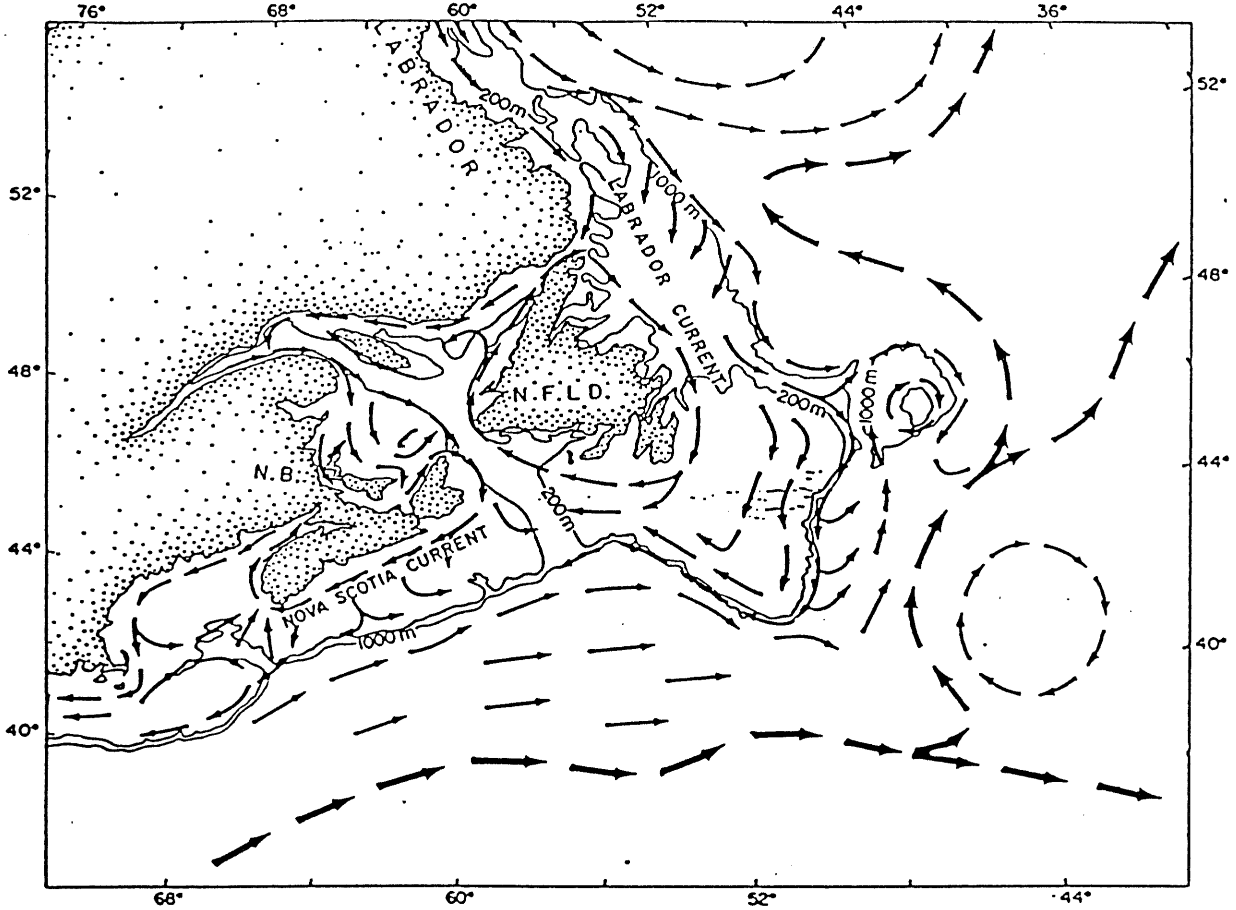


Figure 1.2 Modern surface oceanographic current patterns present in the study area (from Scott *et al.* 1984).

As stated by Marsters (1988), relatively few publications deal with foraminiferal distributions from the Scotian Shelf region. However, recent foraminifera on the Scotian Shelf indicate that assemblage distributions conform closely with local water masses (Williamson 1983). Additional studies include those by Bartlett (1964) on the southeast continental shelf; Barbieri and Medioli (1969) on the Scotian Shelf; Walker (1976) on inshore environments near Halifax; Williamson *et al.* (1984) on the continental margin; and Medioli *et al.* (1986) near Sable Island.

1.5 Organization

Chapter 2 summarizes the methodology of this research, including collection methods, sampling techniques, foraminiferal analysis, and ^{14}C dating procedures. Chapter 3 deals with the results of the foraminiferal analysis and ^{14}C dating. Chapter 4 discusses the results, and Chapter 5 concludes the investigation and summarizes the key points.

CHAPTER 2 METHODS

2.1 Introduction

Piston core 91-018-53P was split, described, photographed, and X-rayed. Foraminifera and mollusc shells provided organic sources for ^{14}C dating. This chapter describes core collection, core sampling and subsequent laboratory procedures.

2.2 Piston Coring

A piston corer is a metal barrel with a plastic inner lining driven into soft sediment by a 1000 kg weight. As shown in Figure 2.1, the mechanism is piston-driven when the pilot core makes contact with the sediment surface. The 75 kg weight in the pilot core drives it into the surface of the sediment. Known as a trigger weight core, it is used in conjunction with the longer piston core because the piston corer may bypass up to 1 m of the uppermost sediment as a result of improper trigger core cable measurement or wave-induced movement of the ship. Sediment bypass occurred with core 91-018-53P, missing the upper 1 m of the core (about 3-4 thousand years). Unfortunately, no trigger weight core was available to compensate for this loss.

The piston, initially located at the bottom of the corer, remains on the sediment surface as the metal barrel and plastic inner lining penetrate the sediment. The action of the piston excludes water from entering the core barrel, assuring that the sediments efficiently enter the plastic liner with minimal compaction and disturbance of the sedimentary layers (Pinet 1992).

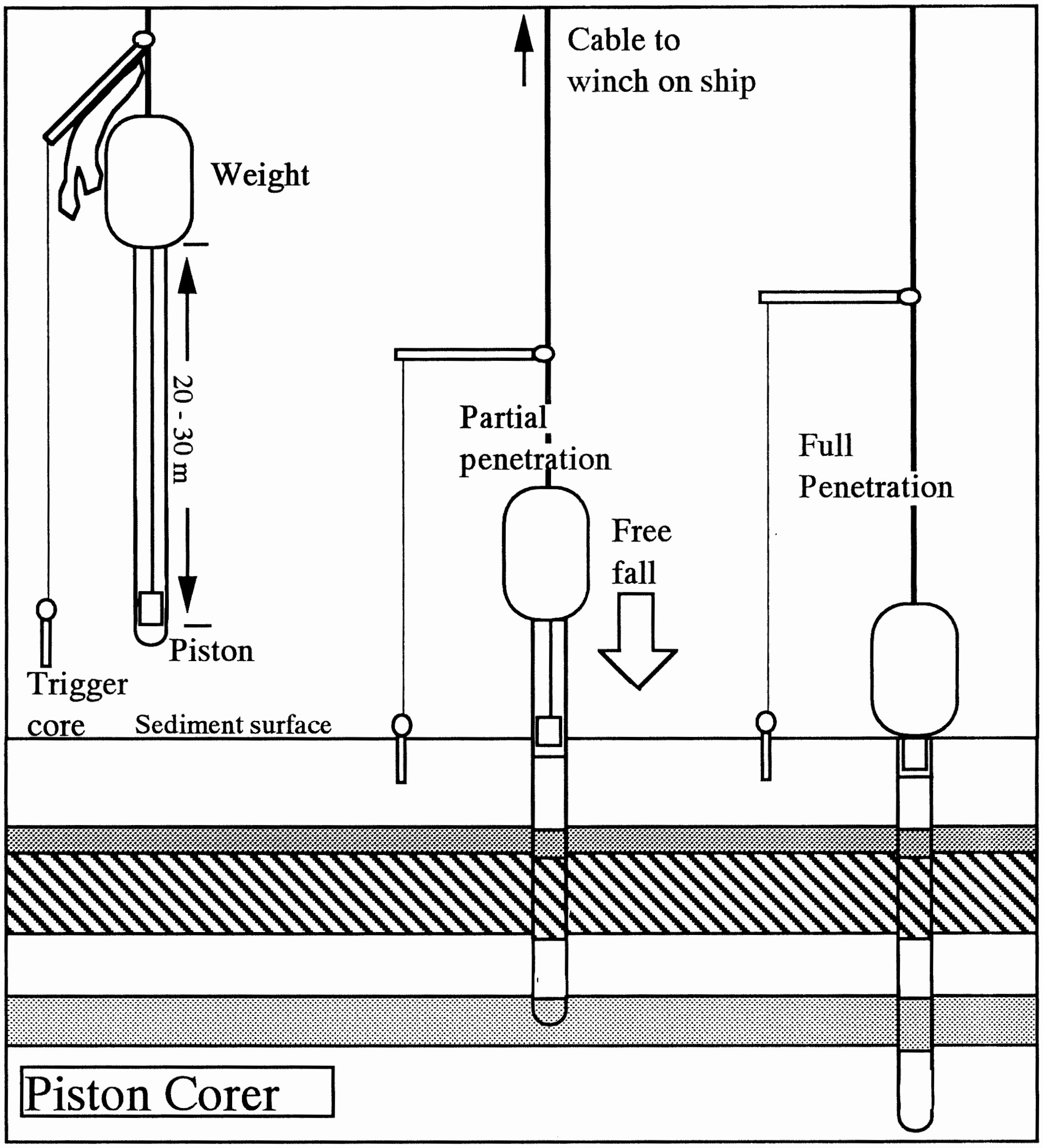


Figure 2.1 Schematic representation of a piston core in operation (after Pinet 1992).

When the action of the piston finishes, the vacuum created beneath the piston retains the sediment as the mechanism is hoisted back to the ship.

The rigid metal barrel used in piston coring routinely obtains cores with lengths less than 15 m, but occasionally more than 20 m (Pinet 1992). The relative simplicity, ease of maintenance, and efficiency of piston coring makes it one of the standard operating procedures for coring into offshore soft sediments. For a more detailed account of piston coring, the reader is recommended to view a video produced by Centre for Marine Geology, Dalhousie University, Halifax, Nova Scotia (Scott 1985).

2.3 Core Analysis

The core was split lengthwise into archival and working halves, photographed, and X-rayed (to highlight structures and stratigraphy otherwise unseen). The X-ray and grain size analysis aided in the core description. Samples (10 cc) were taken at fairly regular intervals (about every 10 cm) in the core for foraminiferal investigation, as well as above and below important sedimentological boundaries. The ^{14}C dating of large shell fragments, as well as some foraminifera, resulted in a chronostratigraphic profile (Fig. 3.6). The core is currently held in cold (2-3^{0}C) storage at Bedford Institute of Oceanography, Dartmouth, Nova Scotia.

2.4 Foraminiferal Processing

The 10 cc samples from the working half of the split core were wet-sieved through a 63 micron sieve. The portions retained by the sieve were collected and stored in a 1:1 solution of water and methanol to minimize bacterial degradation of the sample during storage.

2.5 Foraminiferal Examination

Samples were separated into equal size fractions (i.e., $^1/_8$ or $^1/_6$) using a wet splitter according to the procedures outlined in Scott and Hermelin (1993). These size fractions were examined using a Carl Zeiss binocular microscope (20X, 40X, and 80X magnification) until at least 300 foraminifera were counted; counting statistics require a minimum of 300 individuals to detect species that constitute 1% of the total population with a certainty of 95% (Williamson 1983).

2.6 Carbon-14 Dating

High-energy cosmic particles shattering the nuclei of nitrogen atoms in the upper atmosphere produce carbon-14, an unstable isotope of carbon. This impact releases neutrons that, in turn, produce ^{14}C when they collide with other nitrogen atoms (Fig. 2.2). Production of ^{14}C in the atmosphere is assumed to be in a steady state.

Most of the radiocarbon produced oxidizes to CO_2, some of which enters the hydrosphere. While they are alive, marine organisms acquire amounts of ^{14}C from the water assumed to be the same $^{14}C/^{12}C$ ratio as that in the atmosphere. After they die, no new carbon is added thus, over time, the ^{14}C content decreases by radioactive decay. The rather short ^{14}C half-life of 5730 years (the time it takes for half the ^{14}C present to decay to ^{14}N), together with the sensitivity of counting devices, imposes an upper limit of measurable age

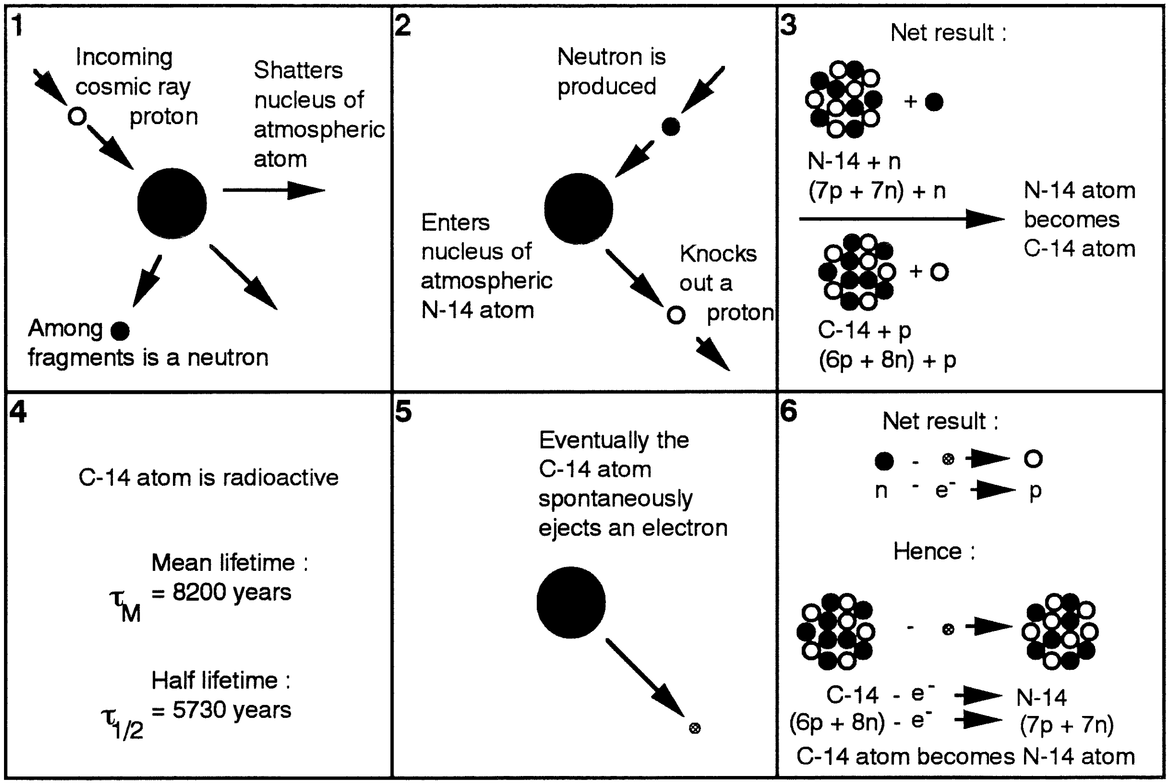


Figure 2.2 Simple physics of atmospheric Carbon-14 generation (after Broecker 1974).

of about 50,000 - 60,000 years. The number of ^{14}C and ^{14}N atoms present in any specimen is so small that direct measurement is difficult. Instead, dating methods use a radiation counter to compare the emission activity (i.e., electron emissions) of a specimen with that of an accurately known standard sample activity. The number of emissions, or 'counts', in a given time period, proportionally relates to the number of atoms of ^{14}C present -- the older the specimen, the less the activity level (Dott and Batten 1988). Recent advances make possible the measurement of ^{14}C in a mass spectrometer such that very small samples of carbonate can be dated. Termed accelerator-mass spectrometer (AMS) ^{14}C dating, it actually measures the isotopic ratios directly (Beukens *et al.* 1986). This technique was used on most of the samples from core 91-018-53P because they were small samples of mollusc shells or foraminiferal tests.

The age limit of 50,000 - 60,000 years means that ^{14}C-dating is useless for most of the geologic timescale but makes it very useful for late Pleistocene and Holocene times. This method of dating is also especially useful in climatology and studies of the movement of groundwater and ocean circulation because determination of the ^{14}C-age of water masses is possible (Dott and Batten 1988). Recently, the discovery of a 'reservoir' effect in marine samples, the time it takes for oceanic waters to recycle, means that some ^{14}C dates in the marine record may be as much as 800 years older than the $^{14}C/^{12}C$ ratios suggest. The amount of reservoir effect depends on the amount of carbon the organism obtained from the open-sea marine environment, determined by ^{13}C measurements separate from the ^{14}C measurements (Beukens *et al.* 1986).

2.7 Summary

Using various methods to analyze a core provides a variety of results and suggests possible reasons for these results. The data obtained from this core are the following : (1) lithological variations by means of a core description involving the use of X-rays, and grain size analysis; (2) paleoceanographic indicators (foraminifera); and (3) chronological sequence determinations (^{14}C ages).

CHAPTER 3 RESULTS

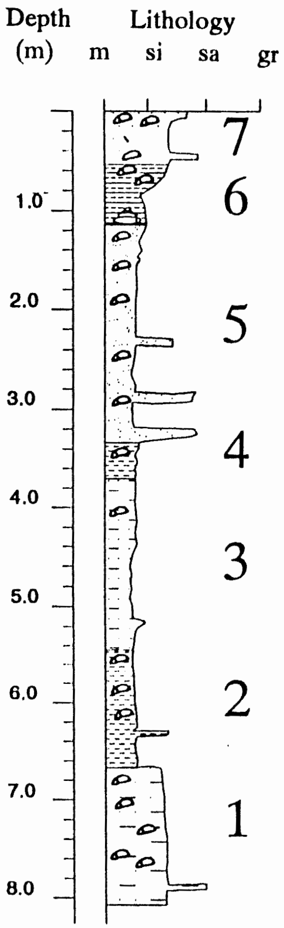
3.1 Introduction

This chapter presents the lithological core description, seismic profile interpretation, foraminiferal assemblages within the core, and ^{14}C ages essential to determining if core 91-018-53P contains foraminiferal evidence of a paleoceanographic change during the Younger Dryas.

3.2 Core Lithology and Seismic Stratigraphy

R. Stea (pers. comm., 1993) described core 91-018-53P as part of his continuing Ph.D. study. Figure 3.1 shows the lithologic profile and brief descriptions of the seven lithofacies present. A Huntec seismic profile of the core site (Fig. 3.2) shows six units separated by unconformities of local and regional extent as well as the approximate core position. The figure caption describes the sedimentary sequence and briefly describes each seismic unit. The top section of the Holocene is not present in core 91-018-53P because of bypassing by the piston core.

The lithologic profile shows alternating olive-grey and dark greyish-brown muds, and the seismic profile displays a similar stratigraphy evident by rhythmic reflections, probably resulting from acoustic impedance contrasts between sand, silt, and massive muds. Red oxidation staining, black mottling, and bioturbation are common throughout the core. The bottom of the core contains silt and sand in the matrix and, based on seismic profiling, is likely a till or massive glaciomarine sediment. These units probably result from



Lithofacies 7 : Olive-grey (5Y 3/2 Munsell Color Chart) massive mud.

Lithofacies 6 : Dark greyish-brown (2.5Y 4/2) mud. Bioturbation at 100 cm.

Lithofacies 5 : Olive-grey (5Y 3/2) massive stiff mud. Bioturbated from 109-240 cm. White silty layer at 230 cm.

Lithofacies 4 : Dark greyish-brown (2.5Y 4/2) mud. Oxidized to reddish hue. Mottling present.

Lithofacies 3 : Olive-grey (5Y 3/2) slightly mottled, bioturbated mud.

Lithofacies 2 : Dark greyish-brown (2.5Y 3/2) mud. Oxidized to reddish hue. Heavy black mottling present.

Lithofacies 1 : Dark greyish-brown (2.5Y 4/2) compacted mud. Red silt and sand in matrix.

Figure 3.1 Lithologic profile of core (Stea, pers. comm., 1993).

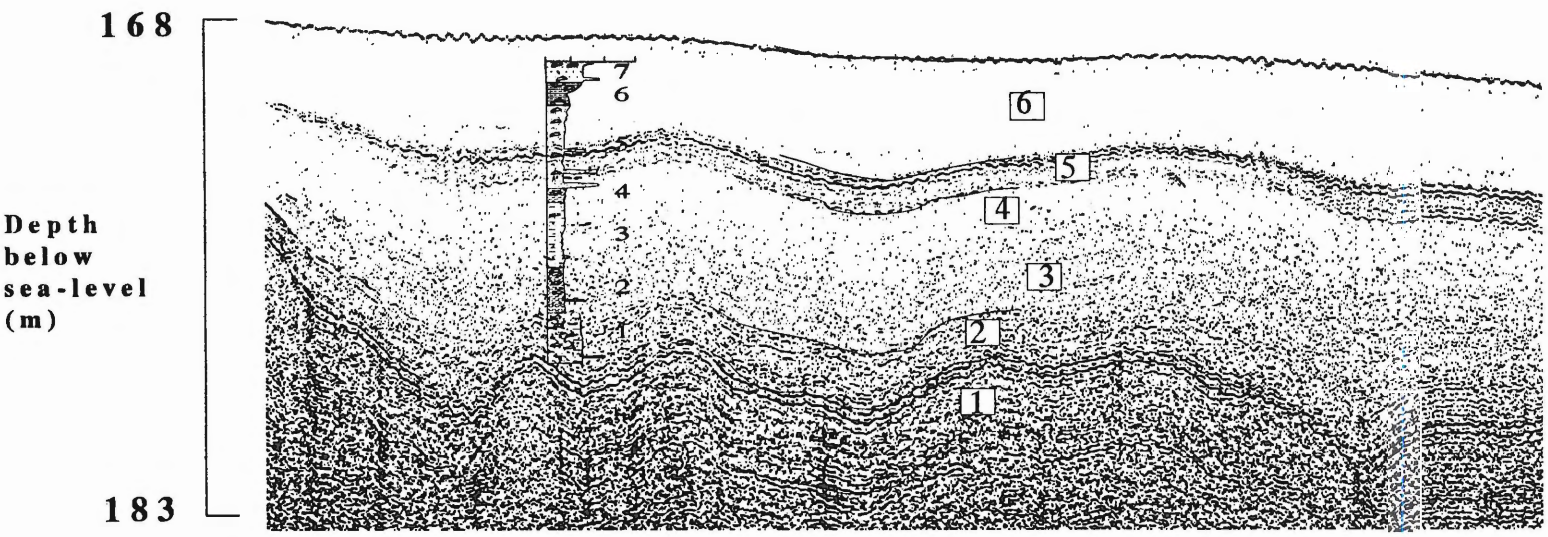


Figure 3.2 Huntec seismic profile showing approximate core position (note that upper Holocene was not sampled with this core). The profile shows six seismic units. Incoherent backscatter and high reflectivity characterize Unit 1, leading to an interpretation as a till or massive glaciomarine sediment capped by a strong reflector. Moderately continuous reflections of low to medium amplitude characterize the draped geometry of Unit 2. In contrast, Units 3 and 4 have lower amplitude reflections and less continuity than Unit 2, and display a more ponded style of deposition. The rhythmic reflections evident in Units 2, 3, and 4 probably result from acoustic impedance contrasts between sand and silt layers and massive muds, leading to their interpretation as proximal and distal glaciomarine sediments. Unit 5 consists of moderate to high amplitude reflections and may reflect a return to proximal glaciomarine conditions. The uppermost unit (6) has low amplitude reflections that onlap onto Unit 5. This represents Holocene marine clay accretion (Stea, pers. comm. 1993).

proximal and distal glaciomarine sediments deposited in response to the repeated advance and retreat of the glacial front.

3.3 Foraminiferal Assemblages

Appendix A shows relative percentages of each foraminiferal species with depth in the core and, from these data, graphs of foraminiferal species were plotted. Lines drawn subjectively throughout the core where at least one foraminiferal species increases or decreases in abundance define foraminiferal assemblages. Because of the greater variety of species of foraminifera in the upper 123 cm of the core, the assemblages from this portion of the core were vertically expanded and graphed separately from the whole core assemblages. Only the dominant species were graphed (eight in total) along with total number of individuals, total number of species, total number of planktonic foraminifera, and total number of ostracods (Figs. 3.3.a and 3.3.b). Figures 3.4.a and 3.4.b show the major assemblages of the whole core.

3.3.1 Assemblage A

Assemblage A ranges from 802 cm to 610 cm depth in the core. Dominant foraminiferal species are: (1) *Elphidium excavatum* forma *clavatum*, a warm ice-margin species (Boyd *et al.* 1988) representing marine ice front conditions; and (2) *Cassidulina reniforme*, a species representing cold waters of glacial or near-glacial domains (Mathewes *et al.* 1993). An assemblage of *E. excavatum* f. *clavatum* and *C. reniforme* characterizes recent and fossil proximal glaciomarine environments in low diversity faunas (Seidenkrantz

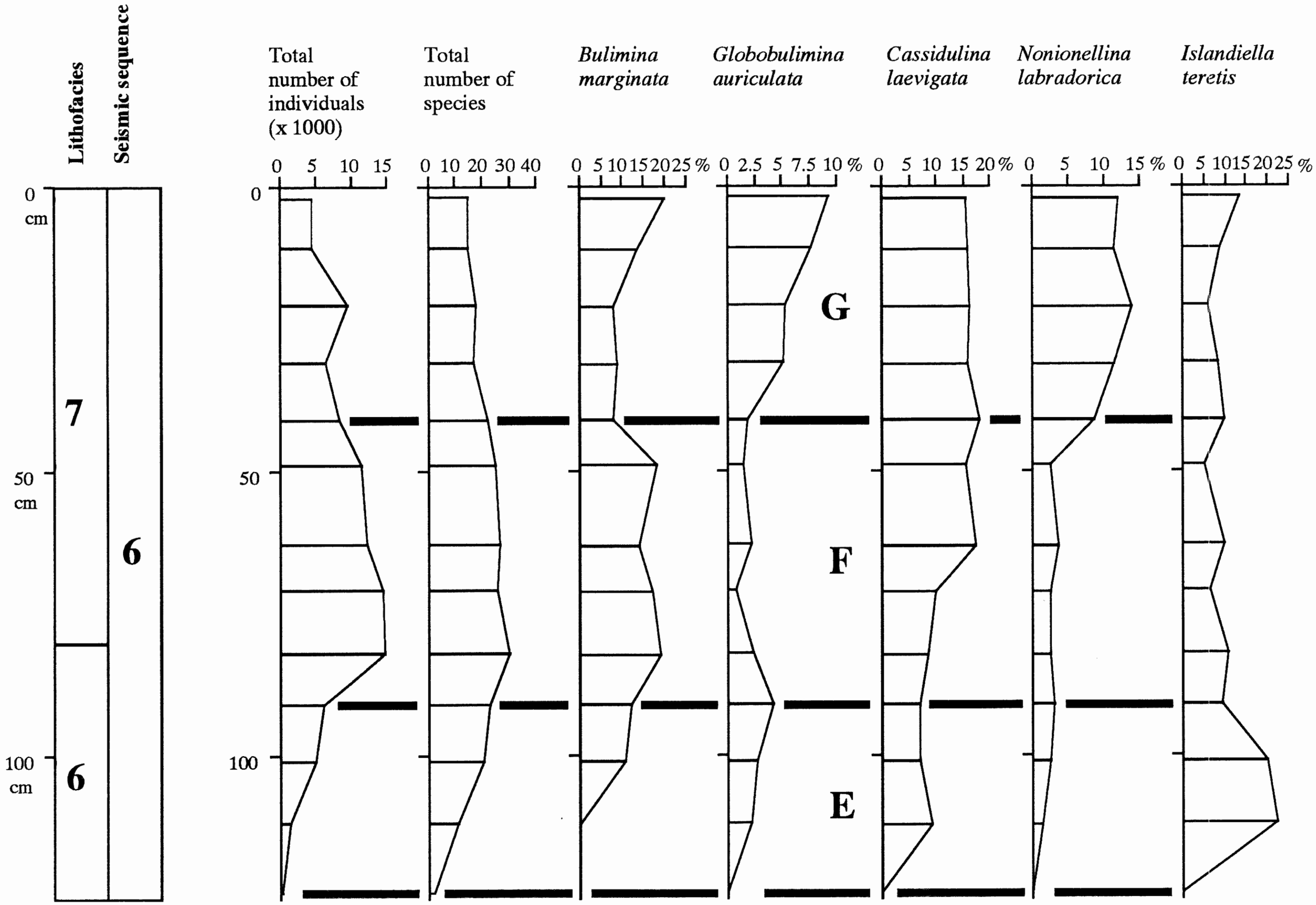


Figure 3.3.a Graphs showing percent foraminifera versus depth, 0-120 cm, core 91-018-53P. E-G are foraminiferal assemblages.

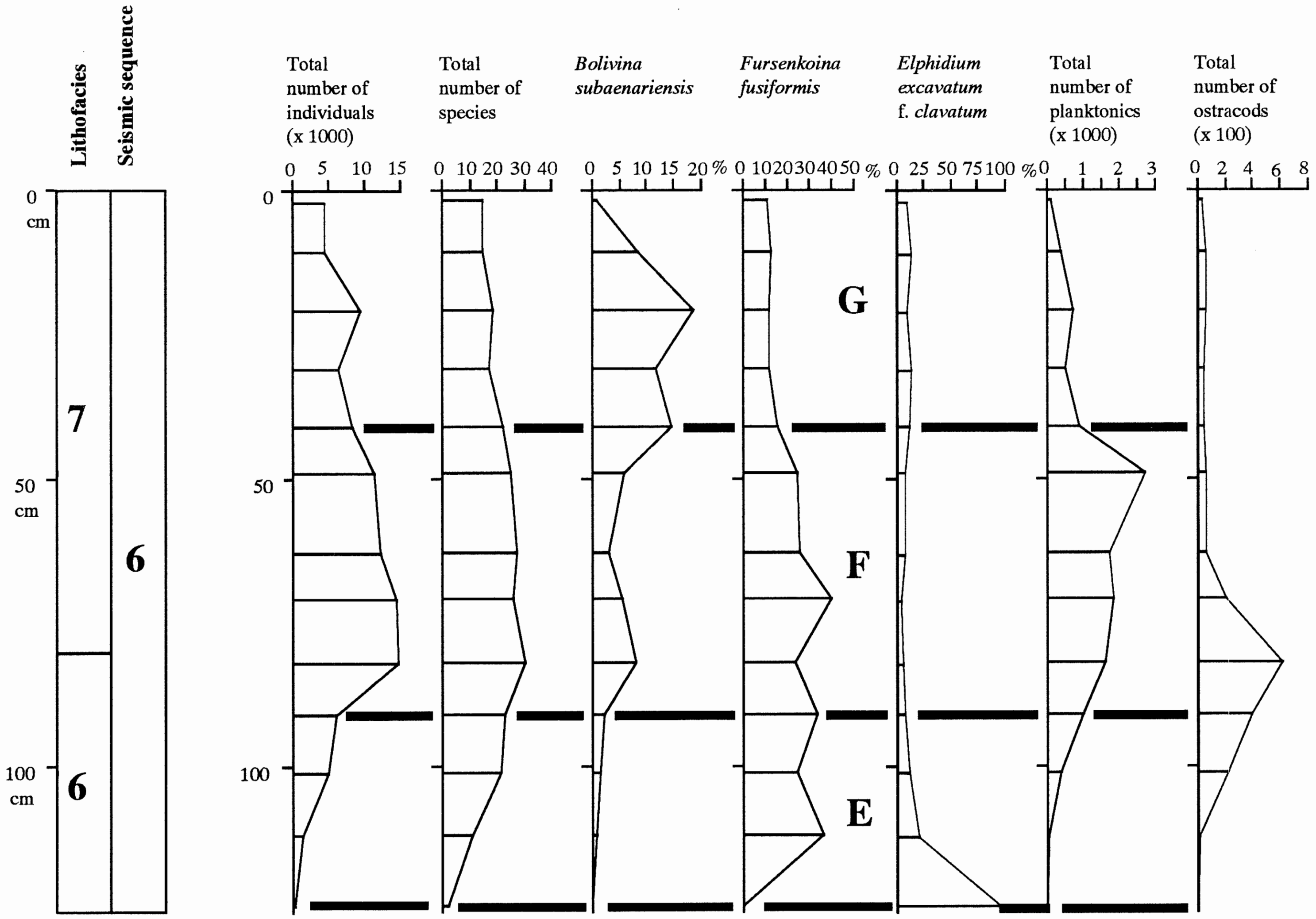


Figure 3.3.b Graphs showing percent foraminifera versus depth, 0-120 cm, core 91-018-53P. E-G are foraminiferal assemblages.

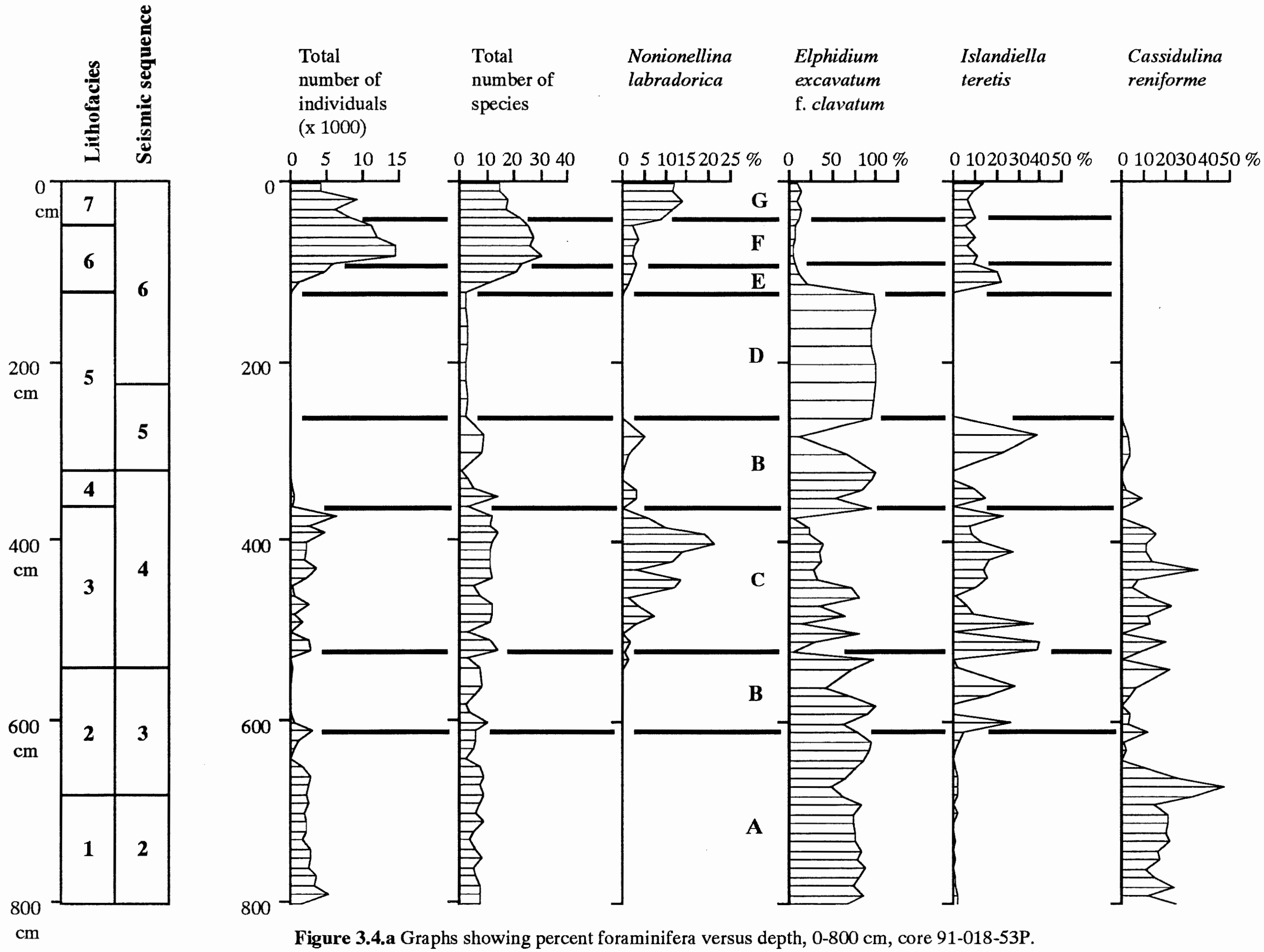


Figure 3.4.a Graphs showing percent foraminifera versus depth, 0-800 cm, core 91-018-53P. A-G are foraminiferal assemblages.

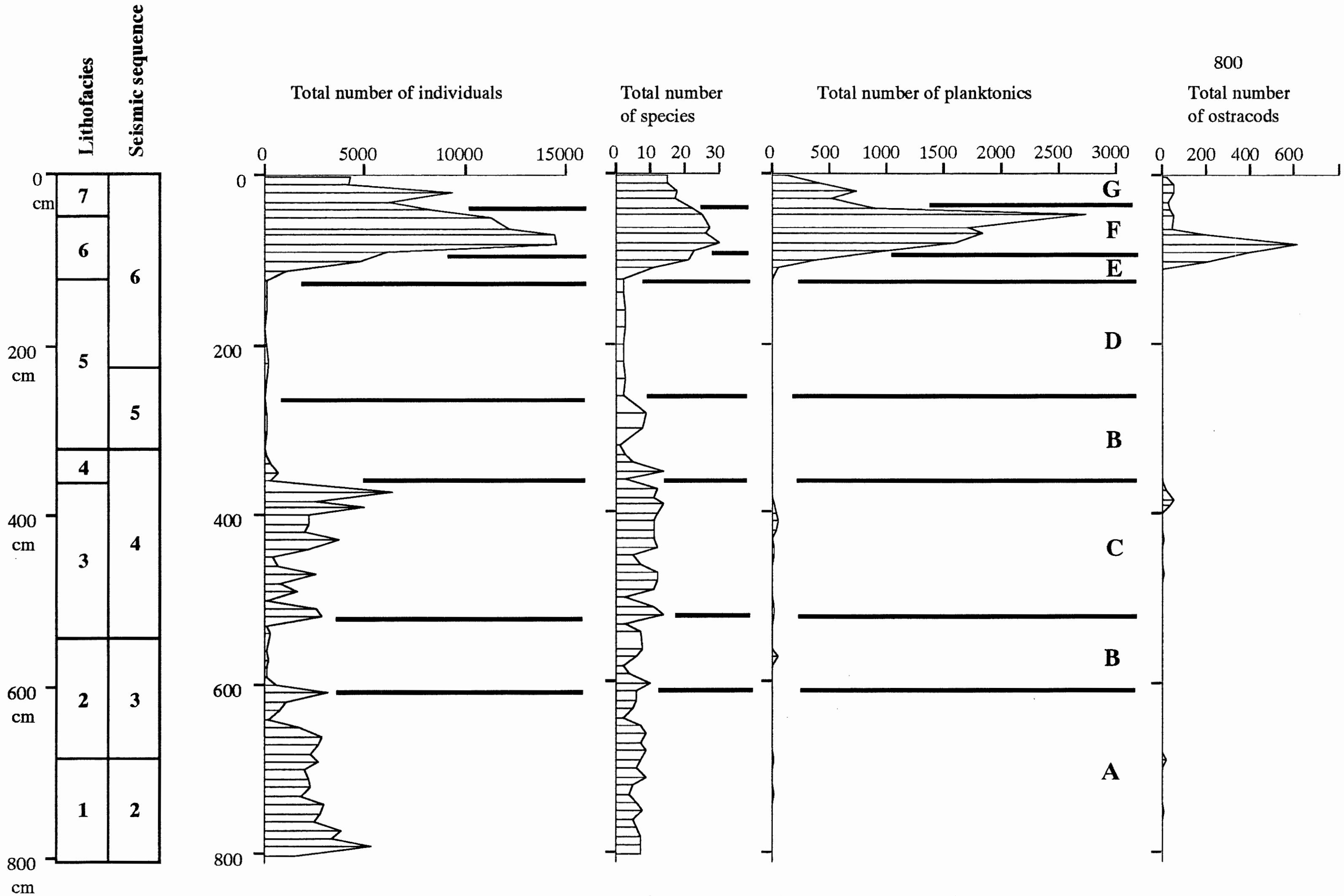


Figure 3.4.b Graphs showing total planktonics and ostracods versus depth, 0-800 cm, core 91-018-53P. A-G are foraminiferal assemblages.

1993). Planktonic foraminifera and ostracods are rare and both total number of foraminifera and foraminiferal species are moderately abundant.

3.3.2 Assemblage B

Assemblage B occurs from 610 cm to 520 cm and recurs at 360 cm to 260 cm depth in the core. Dominant foraminiferal species are: (1) *Nonionellina labradorica*, a species representing warm, saline bottom water (3-4^{0}C, 34.0-34.8‰) which may also represent glaciomarine environments at some distance from the glacier front (Seidenkrantz 1993); (2) *E. excavatum* f. *clavatum*; (3) *Islandiella teretis*, a cool water species; and (4) *C. reniforme*. Planktonic foraminifera are rare and ostracods are absent. The total number of foraminifera is low but the total number of foraminiferal species shows no decrease.

3.3.3 Assemblage C

Assemblage C ranges from 520 cm to 360 cm depth in the core. Dominant foraminiferal species are: (1) *N. labradorica;* (2) *E. excavatum* f. *clavatum*; (3) *I. teretis*; and (4) *C. reniforme*. Planktonic foraminifera and ostracods are rare, but peak approximately coincident with the peak of *N. labradorica* at 401 cm depth in the core (Fig. 3.4.a). Both total number of foraminifera and foraminiferal species are moderately high.

3.3.4 Assemblage D

Assemblage D ranges from 260 cm to 123 cm depth in the core and is almost exclusively *E. excavatum* f. *clavatum*. This assemblage may represent warm ice margin conditions. Total number of foraminiferal species is at a minimum. No planktonic foraminifera or ostracods occur in this assemblage (Fig. 3.4.a). Both total number of foraminifera and foraminiferal species are low.

3.3.5 Assemblage E

Assemblage E ranges from 123 cm to 90 cm depth in the core. Dominant foraminiferal species are: (1) *Bulimina marginata*, a warm water species (Scott *et al.* 1984); (2) *Globobulimina auriculata*, a warm water species (Scott *et al.* 1984); (3) *Cassidulina laevigata*, a warm water species (Scott *et al.* 1984); (4) *N. labradorica*; (5) *I. teretis*; (6) *Bolivina subaenariensis*, a warm water species (Scott *et al.* 1984); (7) *Fursenkoina fusiformis*, a species representing bottom water with low oxygen content (Scott and Medioli 1988); and (8) *E. excavatum* f. *clavatum*. The first major occurrence of planktonic foraminifera and ostracods occur in the upper two thirds of Assemblage E (Figs. 3.3.a and 3.3.b). Both total number of foraminifera and foraminiferal species show an upward increase.

3.3.6 Assemblage F

Assemblage F ranges from 90 cm to 40 cm depth in the core. Dominant foraminiferal species are identical to those of Assemblage E, with marked increases in abundances of *B. marginata*, *C. laevigata*, and *B. subaenariensis*. The total numbers of planktonic foraminifera and ostracods peak at 82 cm depth in the core. Both total number of foraminifera and foraminiferal species peak, indicating warmer environments and greater organic production and diversity (Figs. 3.3.a and 3.3.b).

3.3.7 Assemblage G

Assemblage G ranges from 40 cm to 0 cm depth in the core. Dominant foraminiferal species are identical to those of Assemblage F, with marked increases in abundances of *G. auriculata*, *N. labradorica*, and *B. subaenariensis*, and marked decreases in abundance of *F. fusiformis*. The total numbers of planktonic foraminifera and ostracods decrease rapidly towards the surface of the core, and both total number of foraminifera and foraminiferal species show a similar upward decrease (Figs. 3.3.a and 3.3.b).

3.4 Carbon-14 Dates

Carbon-14 dates from seven calcareous mollusc shells, and two groupings of calcareous foraminifera, produce a line representing time versus depth in the core (Fig. 3.5). The slope of this line is the rate of sedimentation through time. Where the slope between two data points is negative, either an experimental error or reworking took place. As determined by Scott and Medioli (1988), downhole contamination is unlikely because piston coring was the method of core extraction used. Given the turbulent depositional history of the site, reworking probably took place. The effects of this reworking demand that some ^{14}C dates be disregarded when creating the age versus depth curve (Fig. 3.5).

Based on ^{14}C ages, the decreased number of foraminiferal species in Assemblage B from approximately 352 cm to 300 cm depth in the core

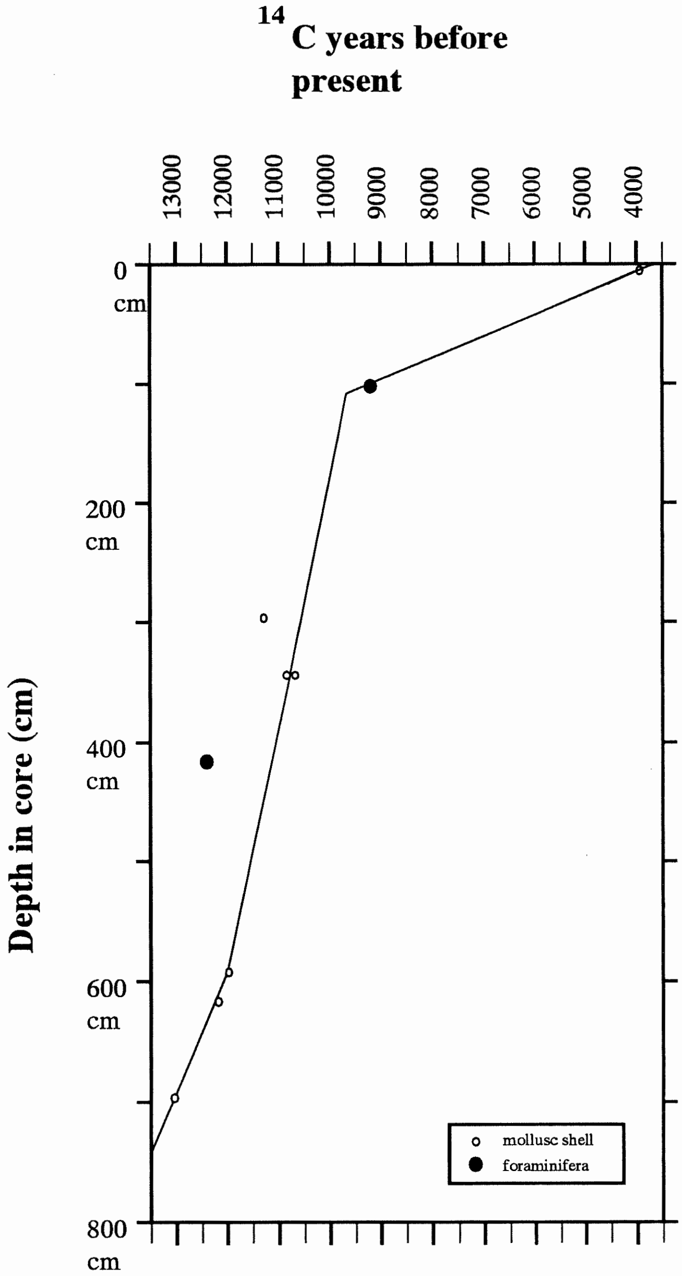


Figure 3.5 Time versus depth in the core based on Carbon-14 dates from seven calcareous mollusc shells and two groupings of calcareous foraminifera. The effects of reworking demand that some dates be disregarded in creating the time versus depth curve.

represents the Killarney Oscillation, a minor cold period pre-dating the Younger Dryas. Assemblage C thus represents a minor warm event post-dating the Killarney Oscillation. Assemblage D represents the Younger Dryas interval and occurs roughly from 260 cm to 123 cm depth in the core. As Figures 3.4.a and 3.4.b show, the Younger Dryas interval contains both a low number of individuals and a low number of species. Reworked ^{14}C dates are not unexpected because there is reworked material in the Younger Dryas section (i.e. diatoms and organic detritus, such as wood fragments). Assemblages E through G represent Holocene warming trends post-dating the Younger Dryas.

3.5 Summary

Lithologic profiles and seismic interpretations (Figs. 3.1 and 3.2, respectively) show the core to be mainly mud and silt with rare sandy layers. The type of sediment and rate of sedimentation are useful in determining the depositional history of an area and, along with foraminiferal assemblages in core 91-018-53P, give insight into activity along the glaciomarine margin, especially important in this paleoceanographic reconstruction.

CHAPTER 4 DISCUSSION

4.1 Introduction

Determining paleoenvironments using foraminiferal assemblages refines and, in some cases, explains sedimentological (i.e. lithofacies) and possibly seismic sequences. This chapter reviews the lithofacies and the foraminifera that characterize them.

A review of the theories for the cause of the Younger Dryas follow the lithofacies discussion. A brief examination of the Younger Dryas event in Nova Scotia attempts to explain similar conditions (foraminiferal and sedimentological) in core 91-018-53P. This chapter concludes with a paleoceanographic reconstruction for the site of core 91-018-53P, and a comparison with previous paleoceanographic models for the surrounding area.

4.2 Intra-Glacial Stratigraphy

Each of the following sections describes one of the seven lithofacies shown in Figure 3.1, and the foraminiferal assemblage that characterizes it. The sections follow the general format of describing the depth and age boundaries (based upon the ^{14}C dates from Figure 3.5), lithology, and foraminiferal assemblages present in the core.

4.2.1 Lithofacies 1

Lithofacies 1, ranging from 802 cm to 665 cm depth in the core, is a glacial deposit extending in this core to around 12,560 ^{14}C ybp. It is a dark greyish-brown compacted mud with some red oxidized silt and sand in the matrix, characterized by the ice margin fauna of foraminiferal Assemblage A. Typically, 5-10 foraminiferal species characterize this lithofacies. Total number of foraminifera in this lithofacies slightly decrease upwards. Trace amounts of both planktonic foraminifera and ostracods also occur.

4.2.2 Lithofacies 2

Lithofacies 2, ranging from 665 cm to 545 cm depth in the core, is a glacial deposit ranging in age from about 12,560 to 11,625 ^{14}C ybp. It is a dark greyish-brown mud with heavy black mottling and some red oxidized areas. In this unit, the lithofacies and foraminiferal assemblages do not perfectly match; it has foraminiferal Assemblage A in the lower third and Assemblage B in the upper two thirds of the unit. Typically, 5-10 foraminiferal species characterize this lithofacies. Total number of foraminifera in this lithofacies decreases upwards to very low amounts. Lithofacies 2 contains only trace amounts of planktonic foraminifera.

4.2.3 Lithofacies 3

Lithofacies 3, ranging from 545 cm to 370 cm depth in the core, is a postglacial deposit ranging in age from about 11,625 to 10,875 ^{14}C ybp. It is an olive-grey and slightly mottled mud characterized by the sudden appearance of Assemblage C, representing cold but non-glacial conditions. Bioturbation indicates the presence of organisms absent in the cooler environment of Lithofacies 2. The number of foraminiferal species is constant throughout

Lithofacies 3, but total number of foraminifera increase upwards with only small amounts of both planktonic foraminifera and ostracods.

4.2.4 Lithofacies 4

Lithofacies 4, ranging from 370 cm to 335 cm depth in the core, is a postglacial deposit ranging in age from about 10,875 to 10,560 ^{14}C ybp. It is a dark grey-brown with minor mottling and red oxidized areas, characterized by Assemblage B. Both the total number of foraminifera and foraminiferal species are low.

4.2.5 Lithofacies 5

Lithofacies 5, ranging from 335 cm to 115 cm depth in the core, is a postglacial deposit ranging in age from about 10,560 to 9,313 ^{14}C ybp. It is a massive stiff mud and olive-grey in color, characterized by Assemblage B in the lower third and Assemblage D in the upper two thirds of this facies. Abundant bioturbation exists in the upper two thirds of the lithofacies. The number of foraminiferal species peaks just below the three major disappearances, and upwardly decreases to about three species at any given point in the upper two thirds of the lithofacies (the Younger Dryas). Total number of foraminifera is the lowest in the core throughout this facies.

4.2.6 Lithofacies 6

Lithofacies 6, ranging from 115 cm to 55 cm depth in the core, is a postglacial deposit ranging in age from about 9,313 to 6,375 ^{14}C ybp. It is a dark greyish-brown mud with some bioturbation, characterized by sharp changes in the foraminiferal assemblages: first Assemblage E in the lower half, and Assemblage F with higher diversity and numbers in the upper half. The highest number of foraminiferal species in the core occurs in this lithofacies, peaking at just under 30/10cc. Additionally, highest abundances of foraminifera occur in this lithofacies, peaking at just under 15,000/10cc. Planktonic foraminifera are abundant, and increase upwards. Ostracods are abundant, with a sharp peak in the lower portion of this lithofacies and a rapid decrease upwards.

4.2.7 Lithofacies 7

Lithofacies 7, ranging from 55 cm to 0 cm depth in the core, is a postglacial deposit beginning around 6,375 ^{14}C ybp. It is a massive olive-grey mud, characterized by the appearance of Assemblage G. Both total number of foraminifera and foraminiferal species gradually decrease upward. Planktonic foraminifera gradually decrease upward and ostracods remain present in moderate numbers throughout the lithofacies.

4.3 Younger Dryas Interval

The Younger Dryas is an event returning to almost full-glacial conditions (Ruddiman and McIntyre 1981; Levesque *et al.* 1993; and Mathewes *et al.* 1993), classically dated from 11,000 to 10,000 ybp. Pollen (Mott *et al.* 1986) and foraminifera (Scott *et al.* 1989a) locally record this climate cooling. Because the glacial advance took place in the Northern Hemisphere, the volumetrically important Laurentian ice sheet plays an important role in the cause and duration of the Younger Dryas. For this reason, most pioneering work on the Younger Dryas took place in northeastern North America and northwestern Europe (Lehman and Keigwin 1992).

4.3.1 Global examples of the Younger Dryas

Although approximate age limits of the Younger Dryas are known, exact dates prove elusive (Cwynar and Watts 1989). In Ballybetagh, Ireland, wood fragments with a date of 10,600 ybp mark the accepted base of the Younger Dryas (Cwynar and Watts 1989). Thus the estimate of 11,000 ybp for the start of the Younger Dryas may not apply to Ireland and suggests that perhaps the Younger Dryas was not a globally synchronous event. The Ballybetagh date is in close agreement with dates from a giant Irish deer bone found immediately below the transition to the Younger Dryas (Barnosky 1985) and terrestrial macrofossils from a small lake on the western Swiss Plateau suggesting dates of 10,350 to 10,900 ybp for the beginning of the Younger Dryas and 9,550 to 9,900 ybp for its end (Cwynar and Watts 1989).

Coral reefs drilled off Barbados provide a continuous and detailed record of sea-level change during the last deglaciation (Fairbanks 1989). Results characterize the Younger Dryas chronozone in the marine record as experiencing minimal glacial meltwater discharge, particularly between 11,000 and 10,500 ybp. These results call for an explanation of the Younger Dryas to be the exact opposite of explanations that call upon maximum influxes of glacial ice and water between 11,000 and 10,000 ybp (Fairbanks 1989). Fairbanks (1989) concluded that regional responses to the meltwater pulses are not globally uniform.

Pollen and chemical stratigraphic studies of lacustrine sediments in the Glacier Bay area of southeastern Alaska suggest that a significant climate reversal occurred between approximately 10,800 and 9,800 ybp (Engstrom *et al.* 1990). This event is similar in timing and severity to the Younger Dryas event in the North Atlantic region and pollen data show shrub- and herb-dominated tundra during this period (Engstrom *et al.* 1990). If this event represents the Younger Dryas, as it likely does, then an explanation for a hemisphere-wide propagation of a North Atlantic climatic perturbation must be sought (Engstrom *et al.* 1990).

Paleoclimatic records from the Pacific coast of Canada indicate a return to a colder climate between about 11,000 and 10,200 ybp. Pollen data record a shift from forest to open, herb-rich vegetation after 11,100 ybp (Mathewes *et al.* 1993). As with the Younger Dryas in Alaska, the presence of the Younger Dryas off the British Columbia coast demands an explanation not only with hemispheric implications, but also with possible global implications (Mathewes *et al.* 1993).

Carbon-14 and ^{230}Th dating of fossil corals from the Huon Peninsula, Papua New Guinea, allowed for calibrating part of the ^{14}C timescale, and for estimating rates of sea-level rise during the last deglaciation (Edwards *et al.* 1993). The record of sea-level rise also shows that average global rates of melting were relatively high at the beginning of the Younger Dryas. The reduced melting after the onset of the Younger Dryas would appear to be a response to cooling during the Younger Dryas, rather than a phenomenon necessarily related to the cause of the Younger Dryas (Edwards *et al.* 1993).

Several lines of evidence exist for a Younger Dryas climatic oscillation in other parts of the world. They include glacial readvances in the Andes (Clapperton 1993), changes in an upwelling record from a marginal Caribbean basin, a rapid change in Ethiopian lake levels at this time, ^{14}C dated records from Tierra del Fuego (Crowley and North 1991), and Antarctic ice cores (Harvey 1989; and Crowley and North 1991). However, some places, such as Tasmania, do not show evidence of the Younger Dryas (Crowley and North 1991) whereas other places, such as New Zealand, show debatable evidence for the Younger Dryas (Harvey 1989; and Crowley and North 1991).

These global examples of the Younger Dryas suggest the need to develop other explanations for the cause of the Younger Dryas. Additionally, As Fairbanks (1989) and Crowley and North (1991) believe the areal extent of effects of the Younger Dryas needs refinement, perhaps leading to a possible re-evaluation of the traditionally-held cause(s) of the Younger Dryas.

4.3.2 Possible causes for the Younger Dryas

Many proposed models for the Younger Dryas exist, but most are complex and contain specific details or proofs not dealt with in this paper. The following is a discussion of a few commonly accepted models considered as causes for the Younger Dryas.

4.3.2.1 Meltwater hypothesis

The meltwater hypothesis proposes that, as a result of ice disintegration in the Northern Hemisphere, pulse-like injections of large volumes of ice and meltwater into the North Atlantic Ocean caused ocean cooling (Ruddiman and

McIntyre 1981; Broecker *et al.* 1988; Crowley and North 1991; and Edwards *et al.* 1993). One suggested cause of these increased volumes of meltwater is an eastward diversion of drainage of the large proglacial lakes on the southern margin of the Laurentide ice sheet (Fig. 4.1), away from the Mississippi River and through the St. Lawrence River into the North Atlantic (Ruddiman and McIntyre 1981; Broecker *et al.* 1988; Crowley and North 1991; and Edwards *et al.* 1993). According to Crowley and North (1991), by about 11,000 ybp the ice margin had retreated sufficiently to allow drainage through the St. Lawrence River system. Outflow of meltwater from these large lakes (especially Lake Agassiz), dammed by the ice sheets, probably caused the low-salinity meltwater influx and subsequent cooling of the North Atlantic (Wright 1989; and Crowley and North 1991). This meltwater influx possibly resulted in a disruption of global thermohaline circulation and the northward transport of heat from the south (Engstrom *et al.* 1990). Evidence from Lake Agassiz clearly indicates that its discharge shifted from the Mississippi River drainage system about 11,000 ybp (Broecker *et al.* 1988).

The other credible meltwater source is tabular iceberg calving from ice shelves, or marine-based portions of ice sheets, in Northern Europe during the Allerød warming period immediately prior to the Younger Dryas (Ruddiman and McIntyre 1981; and Wright 1989). Contours of ice-rafted debris in the North Atlantic indicate an Arctic ice outflow pattern from the disintegration of an ice sheet over the Arctic Ocean and northern Norwegian Sea (Ruddiman and McIntyre 1981) (Fig 4.2).

Additionally, the outflow of cool, low-salinity water into the North Atlantic possibly affected the rate of deep-sea mixing (Shaw 1989; and Crowley

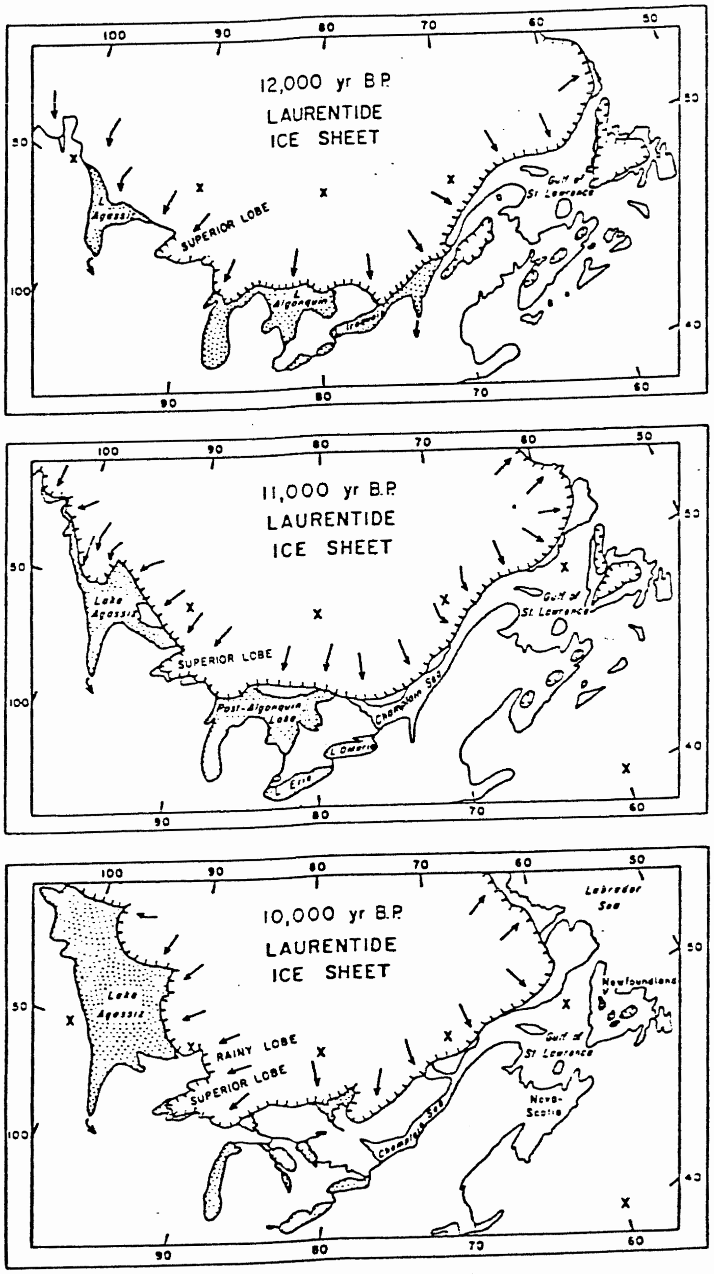


Figure 4.1 Southern margin of the Laurentide ice sheet during glaciation, showing the development of proglacial lakes. Between 11,000 and 10,000 ybp, the Superior and Rainy lobes temporarily retreated and allowed the expanding glacial Lake Agassiz to drain to the Great Lakes and thus to the North Atlantic, perhaps diverting the Gulf Stream, altering the thermohaline circulation patterns in the North Atlantic, and allowing the more southerly drift of icebergs and ice islands from the north (from Wright 1989).

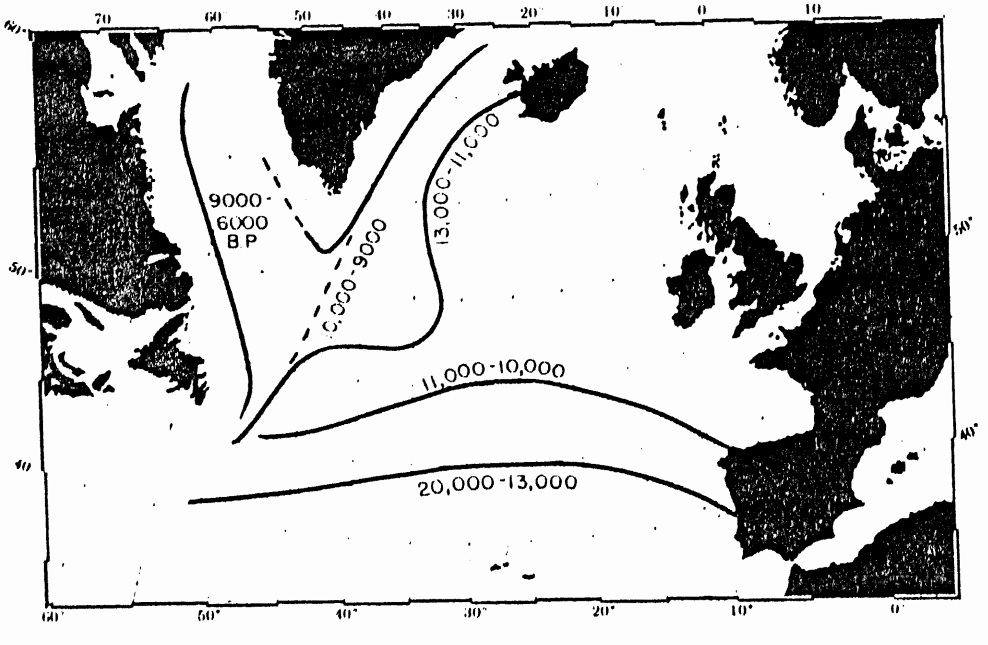


Figure 4.2 Map of glacial retreat and readvances of the North Atlantic polar front based on ice-rafted debris. Though occupied during the major part of intervals indicated, transitions between positions were not necessarily instantaneous (from Ruddiman and McIntyre 1981).

and North 1991). The production of North Atlantic Deep Water (NADW) results in ocean circulation (Fig. 4.3) removing the cold bottom waters to the south and bringing warm surface waters to the north (Crowley and North 1991). A freshwater influx may arrest this deep-sea mixing and result in cooling the waters in the North Atlantic and warming them in the South Atlantic (Broecker *et al.* 1988; Harvey 1989; Shaw 1989; Crowley and North 1991; Lehman and Keigwin 1992; and Sarnthein and Tiedemann 1990).

As Broecker *et al.* (1988) pointed out, the abrupt end of the Younger Dryas may represent a readvance of the Marquette lobe of the Laurentian ice sheet into Lake Superior about 10,000 ybp which once again blocked the eastern outflow of Lake Agassiz. This blockage forced its outflow back into the Mississippi River drainage system, thereby allowing production of NADW and normal deep-sea mixing which, in turn, permits climatic conditions to return to normal (Broecker *et al.* 1988). Greenland ice cores show an extremely abrupt end of the Younger Dryas, perhaps only two decades in duration (Dansgaard *et al.* 1989). The possibility that a lobe of the Laurentian ice sheet could terminate the Younger Dryas supports the hypothesis that such a glacial movement could initiate the Younger Dryas as well.

4.3.2.2 Reorientation of atmospheric flow

The hypothesis of atmospheric flow orientation proposes that the Younger Dryas was a response to a reorientation of westerlies circulation linked to the deglacial changes in ice-sheet extent (Ruddiman and McIntyre 1981). Reorientation of upper atmospheric flow over changing albedo fields at the earth's surface is a valid concept because of an alteration in the pattern of

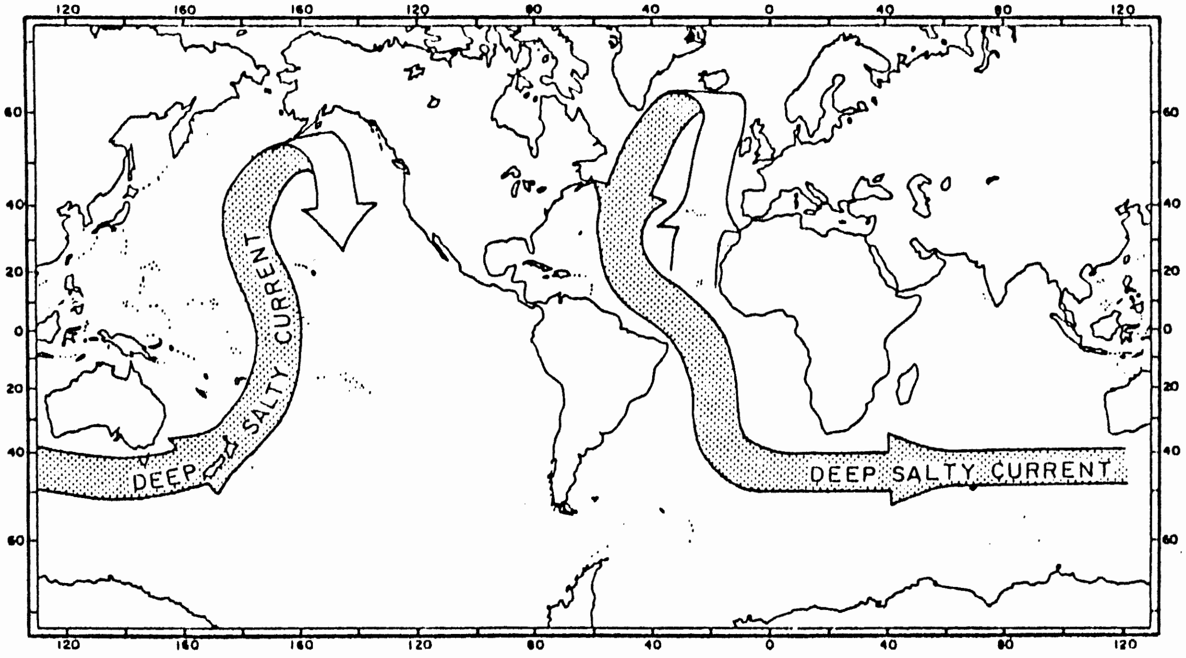


Figure 4.3 A large-scale salt transport current operating in the present-day ocean compensates for the transport of water (as vapor) through the atmosphere from the Atlantic to the Pacific Ocean. Salt-laden deep water formed in the northern Atlantic flows down the length of the Atlantic, around Africa, through the southern Indian Ocean, and finally northward in the deep Pacific Ocean. Some of this water upwells in the northern Pacific, bringing with it the salt left behind in the Atlantic from vapor transport. This coupled atmospheric water vapor transport and ocean salt transport system is self stabilizing. Records from ice and sediment indicate that it was somehow disrupted during glacial times and replaced by an alternative mode of operation (from Broecker *et al.* 1988).

heat flow from the surface. Such a significant shift in flow of the upper westerlies might have different consequences at different longitudes in the circumpolar flow (Ruddiman and McIntyre 1981). However, this change is likely to occur only in the immediate region of the greatest oceanic cooling, and not be global in influence.

Engstrom *et al.* (1990) believe changes in general atmospheric-oceanic circulation may have been an agent for propagating Younger Dryas climatic changes around the globe. Bennett and Boulton (1993) also believe in an instability in the climate-ocean system as a possible cause for the Younger Dryas. Crowley and North (1991) believe that the Younger Dryas may result from meltwater-induced changes in the atmosphere-ocean circulation. The main problem with this explanation is that it cannot account for the short, sharp Younger Dryas event and generally lacks any Younger Dryas analogues in other areas that would be expected by these atmospheric adjustments (Ruddiman and McIntyre 1981).

4.3.2.3 External controls

External controls as the cause of the Younger Dryas represent part of a regular, cyclic response to an external cycle imposed by variations in solar emission (Ruddiman and McIntyre 1981). Another related explanation proposes that long-term changes in incident solar radiation related to Milankovitch cycles caused the Younger Dryas (Lehman and Keigwin 1992). The main problem with this explanation is that the Younger Dryas is much more abrupt than the Milankovitch cycles can explain (Sarnthein and Tiedemann 1990; Lehman and Keigwin 1992; and Bennett and Boulton 1993).

4.3.2.4 Other explanations

The Younger Dryas cooling may not be simply a North Atlantic region phenomenon resulting from the meltwater diversion, ice shelf break-up, arrested deep-water production, reorientation of atmospheric flow, or Milankovitch cycles. Ample evidence exists to suggest worldwide climatic fluctuations corresponding with the Younger Dryas (Mott *et al.* 1986; Harvey 1989; and Crowley and North 1991). Suggestions that the Younger Dryas cooling may be recorded farther inland in continental North America, and elsewhere, suggest that the underlying cause or causes may be more complex than previously believed, and global cooling may be implicated (Mott *et al.* 1986).

4.3.3 Younger Dryas in Nova Scotia

Pollen data from Penhorn, Bluff, and Silver Lakes suggest an open, tundra-like vegetation in Nova Scotia during the Younger Dryas (Ogden 1987) consistent with modern treeless conditions. Conoran, Leak, and Gillis Lakes also show palynological evidence of this tundra-like vegetation during the Younger Dryas (Mott *et al.* 1986). Through lithological, palynological, and chronological data at fourteen sites of buried organic deposits (shallow pond and mire sediments and organic soils that began to accumulate sometime after deglaciation), Mott *et al.* (1986) concluded at about 11,000 ybp, vegetation reverted to fewer trees and more shrubs and herbs in the south, and fewer shrubs and herbs in the north and north-east. This cool episode continued until almost 10,000 ybp when data indicate a climatic warming (Mott *et al.* 1986).

Stratigraphy in a clay quarry in Lantz, N.S., shows a distinct and abrupt climatic change occurred about 10,900 ybp and caused a decline in spruce trees and an increase in scrub birch and other shrubs and herbs (Stea and Mott 1989). Shubenacadie, N.S., shows a decline in spruce about 11,000 ybp and a subsequent increase in abundance of birch, herbs, and grasses (Stea and Mott 1989). Sedimentary sequences throughout Nova Scotia, such as eolian sand deposits commonly produced during glacial episodes, suggest late reactivation of glaciers between 11,000 and 10,000 ybp (Stea and Mott 1989; Forbes 1991; and Stea *et al.* (in press)). An abrupt climatic deterioration at approximately 10,500 ybp is a likely cause for the sudden abandonment of the paleo-Indian occupation site at Debert, N.S. (MacDonald 1968; and Stea and Mott 1989).

4.3.4 Younger Dryas in core 91-018-53P

As shown in both Figures 3.4.a and 3.4.b, a minimum in both total number of foraminifera, and especially total number of foraminiferal species, characterizes the Younger Dryas. Assemblage D, ranging from approximately 260 cm to 123 cm depth in the core, represents the Younger Dryas and is almost exclusively *E. excavatum* f. *clavatum*. According to Figure 3.5, the Younger Dryas in core 91-018-53P is 10,377 to 9,840 ^{14}C ybp. Because of cold meltwater influx, planktonic foraminifera or ostracods are not present in Assemblage D.

The assumed high sedimentation rates at this time (as shown in Fig. 3.5), and the low food supply from the surface waters which created impoverished conditions inhospitable to life, explain the low number of foraminifera and low diversity in foraminiferal species (Crowley and North 1991). King (1969) and Vilks (1981) further confirm the regional glacial influence by stating that dominance of the *Elphidium* species reflects close proximity of continental ice in the marine environment.

However, later work does not support the meltwater hypothesis. Core 83-012-01P (300 m water depth in Emerald Basin on Scotian Shelf) shows virtually identical foraminiferal assemblages as those in core 91-018-53 (Fig. 4.4) but, based on stable oxygen isotopes, shows no indication of a large meltwater event during the Younger Dryas chronozone (Scott *et al.* 1989a). Although Scott *et al.* (1989a) described a sequence that is now recognized as Younger Dryas, core 91-018-53P contains the first well-documented and dated marine sequence on the east coast of North America. The decrease in both total number of foraminifera and total number of foraminiferal species in core 91-018-53P indicates a decrease in organic productivity produced by a cold freshwater influx, yet Wright (1989) suggested no such decrease in foraminiferal abundance from the suspected meltwater influx.

The decreased number of foraminiferal species from approximately 352 cm to 300 cm depth in the core represents what Levesque *et al.* (1993) recently termed the Killarney Oscillation, a minor cold period. Levesque *et al.* (1993) determined the age for this period as 11,160 to 10,910 ^{14}C ybp (subsequently refining the age of the Younger Dryas as 10,770 to 10,000 ^{14}C ybp).

4.4 Paleoceanographic Implications

Carbon-14 ages from Figure 3.5, foraminiferal assemblages from Figures 3.3 and 3.4, and lithologies from Figure 3.1 contribute in forming the basis for the following paleoceanographic reconstruction. Such data suggest that until approximately 12,020 ^{14}C ybp, the Inner Scotian Shelf experienced a high

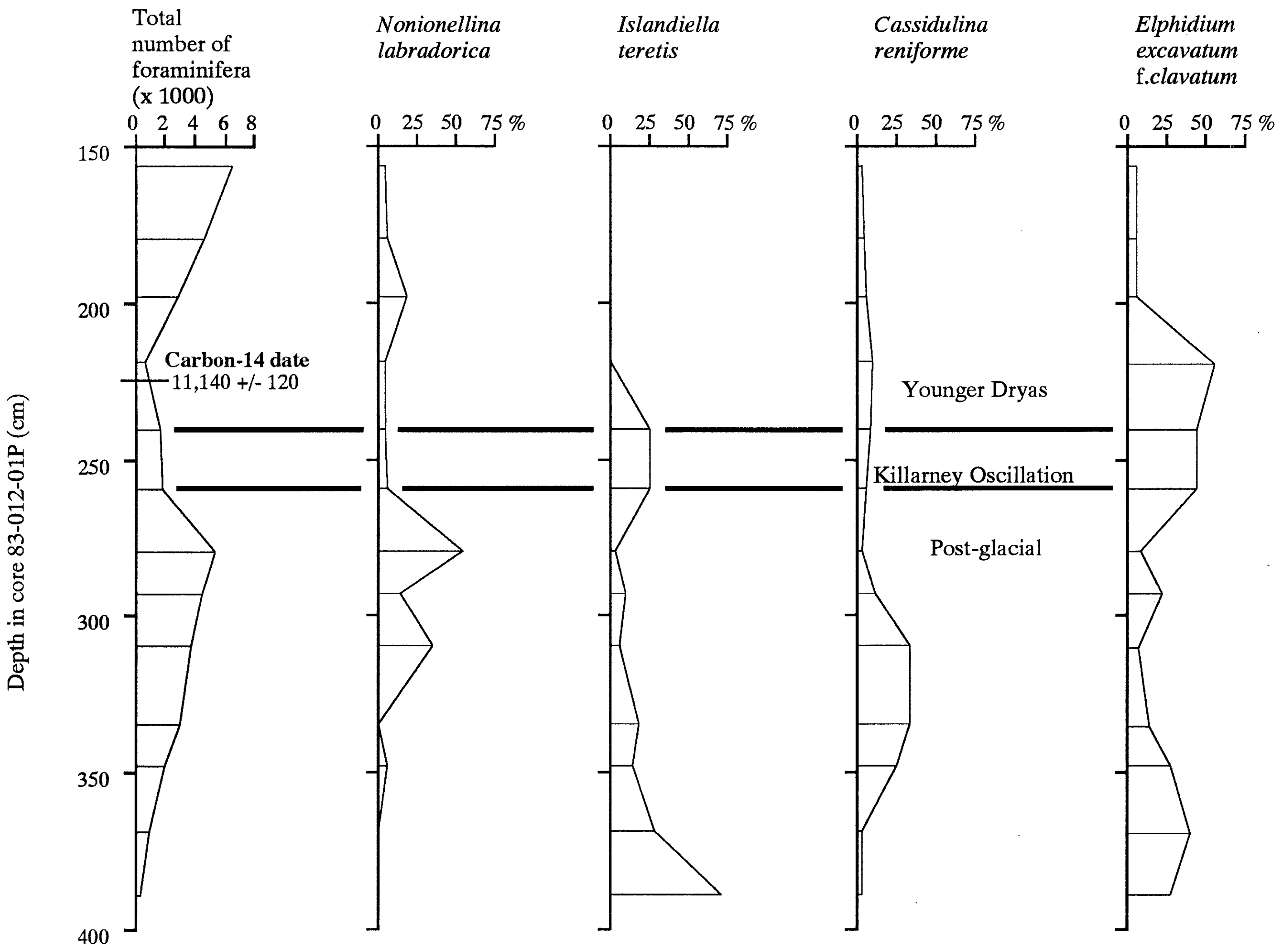


Figure 4.4 Graphs showing percent foraminifera versus depth, 156-389 cm, core 83-012-01P (after Scott *et al.* 1989a).

sedimentation rate (10 cm/century). The dominance of foraminiferal Assemblage A suggests a glaciomarine environment (Scott *et al.* 1984). Seismic reflection profiles support this interpretation by indicating the deposit to be either a till or massive glaciomarine sediment. The high sedimentation rate may have been a result of glacial input along a margin where sea-water temperatures were relatively high (Fairbanks 1989) compared with present-day glacial marine environments such as those in the Canadian Arctic (Scott *et al.* 1984).

Foraminiferal evidence suggests that, although minimized because of extensive glaciation, sea-level appears higher at this time. As a result of marine influences, the *E. excavatum* f. *clavatum* and *C. reniforme* co-dominance decreased with the addition of normal marine species indicated by the appearance of *I. teretis* (Vilks and Rashid 1976). In the time interval 12,020 to 10,805 ^{14}C ybp, sedimentation rates doubled to 20 cm/century and, coupled with the waxing and waning of the glacier front, produced small-scale turbidites represented in the seismic profile as rhythmic alterations between sand and silt layers and massive muds. Biological activity reworked some sediment (suggested by the bioturbation in Lithofacies 3 and the marked increase in total number of foraminifera in Assemblage C). Assemblage C (basically Assemblage B with greater abundance of *N. labradorica*) represents the presence of bottom water of the Outer Labrador Current shortly after deglaciation (Scott *et al.* 1984).

At approximately 10,805 ^{14}C ybp, sedimentation rates increased (23 cm/century) and conditions became warmer and more saline, suggested by the abundance of *N. labradorica* in the upper portion of Assemblage C. An episode represented by Lithofacies 4, which is an oxidized zone similar to the glacial deposit of Lithofacies 2, likely contains the Killarney Oscillation, a minor cool period. Lithofacies 4 almost perfectly aligns with the foraminiferal data showing the recently discovered Killarney Oscillation (Levesque *et al.* 1993), and is the first reported marine occurrence of the Killarney Oscillation. A brief warming episode, characterized by a small increase in *N. labradorica* and a decrease in *E. excavatum* f. *clavatum* follows the Killarney Oscillation.

The upper half of seismic Unit 5, which corresponds to the Younger Dryas chronozone, may reflect a return to proximal glaciomarine conditions. This sedimentation rate (23 cm/century) continues into foraminiferal Assemblage E, a return to warmer water with lower bottom oxygen concentration, suggested by the sudden abundance of *F. fusiformis* in Assemblage E (Scott *et al.* 1984).

Cibicides pseudoungerianus characterizes the slope water, i.e., Atlantic water diluted with about 20% coastal waters (Vilks 1981). This minor foraminiferal species appears at a depth of 102 cm in the core (approximately 9670 ^{14}C ybp.). Above this point, marine clay accretion took place at a rate of about 2 cm/century. Warm, slope-derived waters dominated the core site and continue to do so. Foraminiferal Assemblage G characterizes this zone and continues characterizing present conditions (Vilks and Rashid 1976; Williamson 1983; Scott *et al.* 1984; and Scott and Medioli 1988). This Assemblage G means water with a temperature range of 8-10^{0}C and a salinity of 34‰.

4.5 Comparison with Previous Paleoceanographic Results

As with other studies (Vilks and Rashid 1976; and Vilks 1981), the foraminifera in core 91-018-53P changed from a restricted ice margin fauna in the subsurface layers to the diverse, outer-continental-shelf assemblages in the late Holocene. This faunal change, coinciding with the retreat of ice margins inland, and the establishing of present-day circulation patterns in the North Atlantic (Vilks 1981), occurs at two instances in the core: the base of foraminiferal Assemblage C, and at the base of Assemblage E. These changes may correspond to the two intervals of sea-level fall from 14,000 to 12,000 ybp and, to some degree, from 9,000 to 7,000 ybp as determined by Stea *et al.* (in press). The drops in relative sea-level likely result from isostatic crustal rebound as the ice retreated from the inner shelf.

The erosion and resulting sedimentation during glacial intervals increased measurably when the grounded ice overrode the soft sediments (Scott and Medioli 1988). This erosion suspended large quantities of the underlying sediment which was transported away from the glacial front and deposited seaward (Scott and Medioli 1988).

Scott *et al.* (1984) and Scott *et al.* (1989a) observed benthic foraminiferal results indicating significant paleoceanographic shifts both prior to and during the Holocene. These results, in addition to those of Vilks and Rashid (1976) and Vilks (1981), form the basis for the paleoceanographic reconstruction from core 91-018-53P.

4.6 Summary

The Younger Dryas in this core (foraminiferal Assemblage D) corresponds well with, and provides similar dates (10,377 to 9,840 ^{14}C ybp) to, land (pollen) data for the same interval. That the age of the Younger Dryas in core 91-018-53P does not correspond exactly with the classically-accepted age of 11,000 to 10,000 ybp is probably a result of imprecision between ^{14}C dates (as a result of reworking), not a real age difference. The recently discovered Killarney Oscillation, represented as seismic Unit 4, appears in the lower portion of the upper foraminiferal Assemblage B.

The foraminiferal evidence in core 91-018-53P supports the meltwater hypothesis as a leading cause for the Younger Dryas. The near total dominance of *E. excavatum* f. *clavatum* in this interval represents cooling of the waters and both the high sedimentation rate (23 cm/century), and extensive reworked material (diatoms and wood fragments) represents strong current activity. Figure 4.5 summarizes critical chronostratigraphic correlations among foraminiferal assemblages, lithofacies, and seismic sequences.

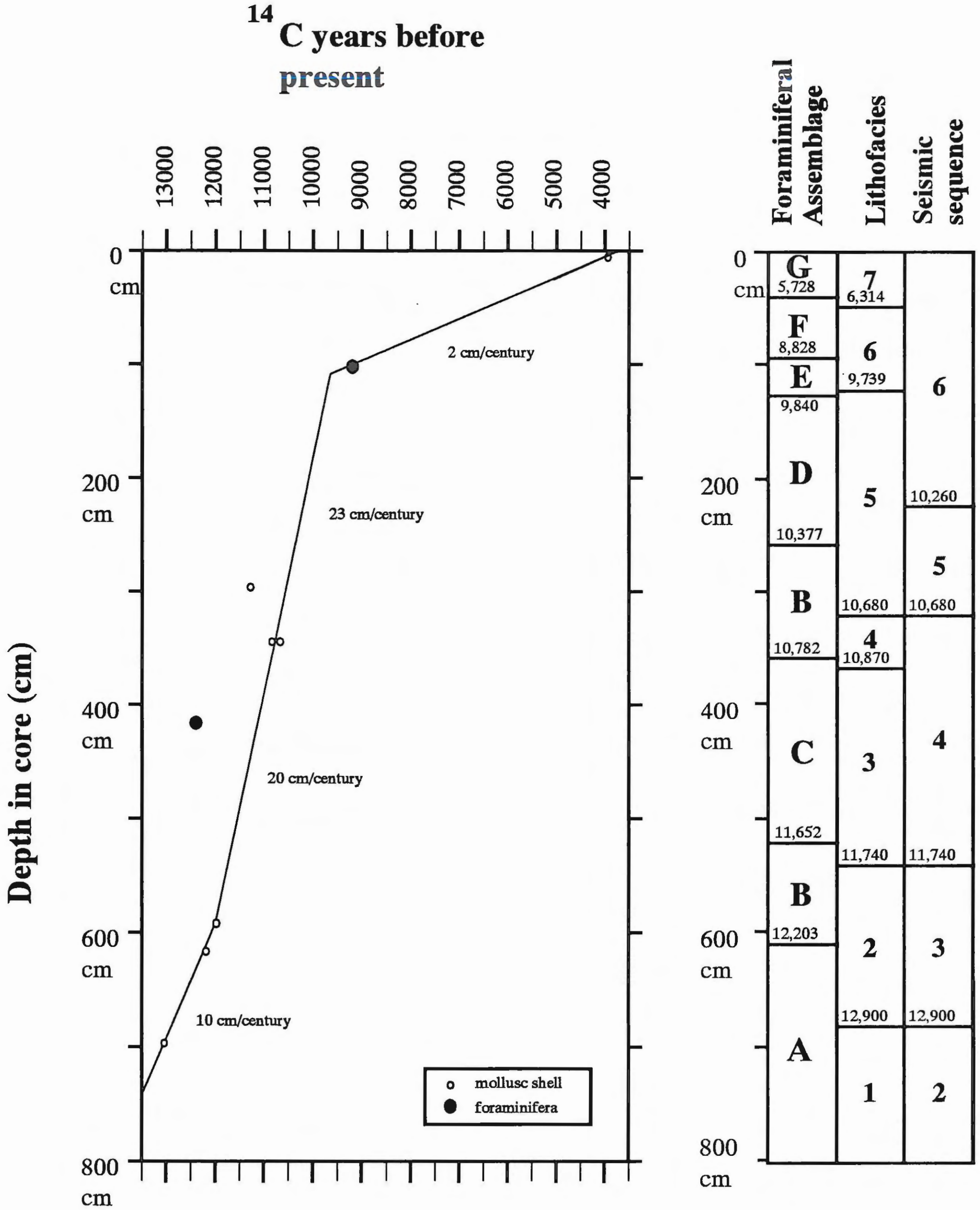


Figure 4.5 Summary diagram of core 91-018-53P, showing rates of sedimentation and the correlation among foraminiferal assemblages, lithofacies, and seismic sequences.

CHAPTER 5 CONCLUSIONS

5.1 Conclusions

The occurrences of assemblages of benthic foraminifera are environmentally-dependent, so changes in fossil assemblages can indicate paleoecologic and paleoceanographic changes. Evidence for reworking in the core (abundant diatoms and wood fragments) demands that some ^{14}C dates be disregarded in creating the chronostratigraphic profile of the core. Additionally, a sharp increase in the number of foraminiferal species at the onset of the Holocene suggests a general, overall warming trend.

In relatively shallow areas, such as continental shelves, bottom water changes generally reflect surface conditions. As a result, benthic foraminiferal evidence supporting the meltwater hypothesis as the cause of the Younger Dryas applies to the whole water column, not exclusively the bottom water environment. Carbon-14 dates indicate that sedimentation rates were five to ten times higher than current rates.

The Younger Dryas representation in core 91-018-53P is the first well-documented marine occurrence on the North Atlantic margin. Additionally, it has a clear (expanded) signal corresponding to a seismic sequence allowing for increased accuracy in regional offshore correlation of the Younger Dryas seismically, lithologically, and foraminiferally. Compared with previous work on the Younger Dryas, results from core 91-018-53P reinforce the notion that the areal extent of the Younger Dryas needs refinement, perhaps leading to a possible re-evaluation of the traditionally-held cause(s) of the Younger Dryas. This core also shows the first reported marine occurrence of the recently discovered Killarney Oscillation, a minor cold period first established by palynological evidence from lakes in Nova Scotia and New Brunswick.

Generally, the good correlation among lithofacies, seismic sequences, and foraminiferal assemblages allows for a paleoceanographic reconstruction along the glaciomarine margin with lithologic and seismic constraints.

SYSTEMATIC TAXONOMY

Astracolus hyalacrulus (Loeblich and Tappan)
Astracolus hyalacrulus LOEBLICH and TAPPAN 1953, p. 52, pl. 9, fig. 1-4.- COLE 1981, p. 58, pl. 17, fig. 48.

Bolivina subaenariensis (Cushman)
Bolivina subaenariensis CUSHMAN 1922, p. 46, pl. 7, fig. 6.- PHLEGER and PARKER 1951, p. 15, pl. 7, figs. 8-10.- COLE and FERGUSON 1975, p. 33.- WILLIAMSON 1983, p. 239, pl. 3, fig. 11.

Brizalina pseudopunctata (Hoeglund)
Bolivina pseudopunctata HOEGLUND 1947, p. 273, pl. 24, fig. 5, pl. 32, figs. 23, 24.- COLE 1981, p. 87, pl. 11, fig. 12.

Buccella frigida (Cushman)
Pulvinulina frigida CUSHMAN 1921 (1922), p. 144.
Eponides frigida (Cushman) var. *calida* CUSHMAN and COLE 1930, p. 98, pl. 13, figs. 13a-c.- PHLEGER and WALTON 1950, p. 277, pl. 2, fig. 21.- PARKER 1952b, p. 449, pl. 5, figs. 3a, b.
Epionides frigidus (Cushman).- CUSHMAN 1942, p. 37, pl.9, figs. 16, 17.- PARKER 1952b, p. 449, pl. 5, figs. 2a, b.
Buccella frigida (Cushman).- ANDERSEN 1952, p. 144, figs. 4a-c, 5, 6a-c.- VILKS 1969, p. 49, pl. 3, fig. 7.- GREGORY 1971, p. 220, pl. 12, figs. 1-3.- COLE and FERGUSON 1975, p. 33, pl. 8, figs. 8,9.- WILLIAMSON 1983, p. 241, pl. 4, fig. 4, 5.

Bulimina marginata (d'Orbigny)
Bulimina marginata D'ORBIGNY 1826, p. 269, pl. 12, figs. 10, 12.- PHLEGER and PARKER 1951, p. 16, pl. 7, fig. 27-28.- FEYLING-HANSSEN 1964, p. 303, pl. 14, fig. 2-3.- SCHNITKER 1971, p. 194, pl. 5, fig. 5.

Cassidulina laevigata (d'Orbigny)
Cassidulina laevigata D'ORBIGNY 1826, p. 282, pl. 15, figs. 4, 5.- SCOTT 1987, p. 377, pl. 2, fig. 10.

Cassidulina reniforme (Norvang)
Cassidulina crassa var. *reniforme* NORVANG 1945, p. 41, text-figures 6c-h.
Cassidulina crassa (d'Orbigny).- SCOTT ET AL. 1980, p. 226, pl. 4, figs. 1-2.
Cassidulina reniforme (Norvang).- SEJRUP and GUILBAULT 1980, pp. 79-85, text-figures 1-3.

Chilostomella oolina (Schwager)
Chilostomella oolina SCHWAGER 1877, p. 10, pl. 1, fig. 16.

Cibicides pseudoungerianus (Cushman)
Truncatulina pseudoungerianus CUSHMAN 1922, p. 97, pl. 20, fig. 9.
Cibicides pseudoungerianus (Cushman).- FEYLING-HANSSEN 1964, p. 340, pl. 19, fig. 4-6.- SCHAFER and COLE 1978, p. 22.- SCHNITKER 1971, p. 196, pl. 9, fig.7.

Cribrostomoides jeffreysi (Williamson)
Nonionina jeffreysi WILLIAMSON 1858, p. 34, pl. 3, figs. 72, 73.
Cribrostomoides jeffreysi (Williamson).- BARBIERI and MEDIOLI 1969, p. 855, fig. 4.- VILKS 1969, p. 45, pl. 1, figs. 17a, b.- COLE 1981, p.30, pl.6, fig.6.

Cyclogyra involvens (Reuss)
Operculina involvens REUSS 1850, v. 2, p. 370, pl. 46, fig. 30.

Dentalina **spp.**

Eggerella advena (Cushman)
Verneuilina advena CUSHMAN 1921, p. 141.
Eggerella advena (Cushman).- CUSHMAN 1937, p. 51, pl. 5, figs. 12-15.- SCOTT and MEDIOLI 1980, p. 38, pl. 2, fig. 7.

Elphidium excavatum (Terquem) **forma *clavatum*** Cushman
Elphidium incertum (Williamson) var. *clavatum* CUSHMAN 1930, p. 20, pl. 7, fig. 10.
Elphidium incertum (Williamson) and variants PARKER 1952a, p. 448, pl. 3, fig. 16.
Elphidium excavatum (Terquem) forma *clavatum* MILLER ET AL. 1982, p. 124, pl. 1, figs. 5-8; pl. 2, figs. 3-8; pl. 4, figs. 1-6; pl. 5, figs. 4-8; pl. 6, figs. 1-5.

Fissurina marginata (Montagu)
Vermiculum marginatum MONTAGU 1803, p. 524.
Lagena sulcata marginata (Montagu).- PARKER and JONES 1865, p. 355, pl. 13, figs. 42, 43.
Lagena marginata (Walker and Jacob).- BRADY 1884, p. 476, pl. 59, fig. 22.
Entosolenia marginata (Montagu?).- CUSHMAN 1948, p. 65, pl. 7, fig. 7.
Fissurina marginata (Montagu).- LOEBLICH and TAPPAN 1953, p. 77, pl. 14, figs. 6-9.- VILKS 1969, p. 48, pl. 2, fig. 2a, b.- GREGORY 1971, p. 207, pl. 10, fig. 1, COLE 1981, p. 81, pl. 19, fig. 25.- WILLIAMSON 1983, p. 239, pl. 3, fig. 8.

Fursenkoina fusiformis (Williamson)
Bulimina pupoides (d'Orbigny) var. *fusiformis* WILLIAMSON 1858, p. 64, pl. 5, figs. 129, 130.
Bulimina fusiformis (Williamson).- HOEGLUND 1947, p. 232, pl. 20, fig. 3, text-figures 219-233.
Virgulina fusiformis (Williamson).- FEYLING-HANSSEN 1964, p. 307, pl. 14, figs. 15-18.
Fursenkoina fusiformis (Williamson).- GREGORY 1971, p. 232,.- SCOTT 1977, p. 172, pl. 7, figs. 9, 10.- WILLIAMSON 1983, p. 226, pl. 5, fig. 1.

Globobulimina auriculata (Bailey)
Bulimina auriculata BAILEY 1851, p. 12, pl. 1, figs. 25-27.

Gyroidina neosoldanii (d'Orbigny)
Rotalia (Gyroidina) *soldanii* d'ORBIGNY 1826, p. 278, no. 5, mod. 36.
Gyroidina neosoldanii BROTZEN 1936, p. 158.

Hoeglundina elegans (d'Orbigny)
Rotalia elegans D'ORBIGNY 1826, p. 276, no. 54.
Epistomina elegans (d'Orbigny).- CUSHMAN 1931, p. 65, pl. 13, fig. 6.
Hoeglundina elegans (d'Orbigny).- COLE 1981, p. 116, pl. 14, fig. 10.- WILLIAMSON 1983, p, 243, pl. 5, figs. 11, 12.

Islandiella teretis (Tappan)
Cassidulina teretis TAPPAN 1951, p. 7, pl. 1, figs. 30a-c.
Islandiella teretis (Tappan).- VILKS 1969, p. 49, pl. 3, fig. 5.- WILLIAMSON 1983, p. 239, pl. 3, fig. 19.

***Lagena* spp.**

Lenticulina peregrina (Schwager)
Cristellaria peregrina SCHWAGER 1866, p. 245, pl. 7, fig. 89.

Nonionella turgida (Williamson)
Rotalina turgida WILLIAMSON 1858, p. 50, pl. 4, figs. 95-97.
Nonionina turgida (Williamson).- BRADY 1884, p. 731, pl. 109, figs. 17-19.
Nonionella turgida (Williamson).- CUSHMAN 1930, p. 15, pl. 6, figs. 1-4.- WILLIAMSON 1983, p. 228, pl. 4, fig. 13.

Nonionellina labradorica (Dawson)
Nonionina labradorica DAWSON 1860, p. 191, fig. 4.
Nonion labradorica (Dawson).- CUSHMAN 1927, p. 148, pl. 2, figs. 7-8.
Nonionellina labradorica (Dawson).- SCHAFER and COLE 1978, p. 28, pl. 9, fig. 4.- COLE 1981, p. 110, pl. 13, fig. 20.- WILLIAMSON 1983, p. 241, pl. 4, fig. 14, 15.

Pullenia subcarinata (d'Orbigny)
Nonionina subcarinata D'ORBIGNY 1839, p. 28, pl. 5, figs. 23, 24.

Quinqueloculina agglutinata (Cushman)
Quinqueloculina agglutinata CUSHMAN 1917, p. 43, pl. 9, fig. 2.- CUSHMAN and TODD 1947, p. 61, pl. 14, figs. 12, 13.- LOEBLICH and TAPPAN 1953, p. 39, pl. 15, figs. 1-4.

Quinqueloculina seminulum (Linne)
Serpula seminulum LINNE 1758, p. 786.
Quinqueloculina seminulum (Linne).- D'ORBIGNY 1826, p. 303.- GREGORY 1971, p. 187, pl. 6, fig. 1.- COLE and FERGUSON 1975, p. 40, pl. 10, fig. 7.- SCOTT 1977, p. 175, pl. 7, figs. 3-5.- SCHAFER and COLE 1978, p. 29, pl. 12, fig.4.
Miliolina seminulum (Linne).- WILLIAMSON 1858, p. 85, pl. 7. figs. 183-185.
Quinqeuloculina seminula (Linne).- CUSHMAN 1929, p. 59, pl. 9, figs. 16-18,.- PARKER 1952a, p. 406, pl.3, figs. 21a, b, 22a, b, pl. 4, figs. 1, 2.- PARKER 1952a, p. 456, pl.2, figs. 7a, b.

Recurvoides turbinatus (Brady)
Haplophragmium turbinatus BRADY 1881a, p. 50.
Recurvoides turbinatus (Brady).- PARKER 1952, p. 402, pl. 2, figs. 23, 24.- VILKS 1969, p. 45, pl. 1, fig. 19.- GREGORY 1971, p. 176, pl. 3, figs. 3, 4.- COLE 1981, p. 32, pl. 6, figs. 7, 8.

Reophax bacillaris (Brady)
Reophax bacillaris BRADY 1881b, p. 49.- CUSHMAN 1910, p. 86, fig. 120.- BARKER 1960, pl. 30, figs. 23, 24.- SCHROEDER 1986, p. 42, pl. 16, figs. 1, 2.- SCHROEDER ET AL 1988, p. 35, pl. 5, fig. 3.

Reophax scorpiurus (Montfort)
Reophax scorpiurus MONTFORT 1808, p. 330.- LOEBLICH and TAPPAN 1953, p. 24, pl.2, figs. 7-10.- LESLIE 1965, p. 169, pl. 1, figs. 6, 7.

Reophax scottii (Chaster)
Reophax scottii CHASTER 1892, p. 57, pl. 1, fig. 1.- WILLIAMSON 1983, p. 207, pl. 1, fig. 11.

Saccammina difflugiformis (Brady)
Reophax difflugiformis BRADY 1879, p. 51, pl. 4, figs, 3a, b.
Proteonina difflugiformis (Brady).- PHLEGER and PARKER 1955, p. 2, pl. 1, figs. 4, 5.
Saccamina difflugiformis (Brady).- TODD and BRONNIMAN 1957, p. 52, pl. 1, fig. 15.- WILLIAMSON 1983, p. 235, pl. 1. fig. 3.

Siphotextularia rolshauseni (Phleger and Parker)
Siphtextularia rolshauseni PHLEGER and PARKER 1951, p. 4, pl. 1, figs. 23, 24a, b.- PHLEGER ET AL 1953, p. 26, pl. 5, fig. 7.- COLE 1981, p. 36, pl. 5, fig. 7.

Spiroplectammina biformis (Parker and Jones)
Textularia agglutinans (d'Orbigny) var. *biformis* PARKER and JONES 1865, p. 370, pl. 15, figs. 23, 24.
Spiroplecta biformis (Parker and Jones).- BRADY 1878, p. 376, pl. 45, figs. 25-27.
Spiroplectammina biformis (Parker and Jones).- CUSHMAN 1927, p. 23, pl. 5, fig. 1.- PARKER 1952a, p. 402, pl. 3, figs. 1, 2.- GREGORY 1971, p. 177, pl. 3, fig. 6.- COLE and FERGUSON 1975, p. 42, pl. 3, fig. 3.- SCOTT 1977, p. 176, pl. 3, fig. 4.- SCHAFER and COLE 1978, p. 19, pl3, fig. 2.

Trifarina angulosa (Williamson)
Uvigerina angulosa WILLIAMSON 1858, p. 67, pl. 5, fig. 140.
Angulogerina angulosa (Williamson).- LESLIE 1965, p. 155, pl. 18, fig. 13a-c.
Trifarina angulosa (Williamson).- BARBIERI and MEDIOLI 1969, p. 857, pl. 65, fig, 1-5.- COLE 1981, p. 92, pl. 19, fig, 46.- WILLIAMSON 1983, p. 241, pl. 4, fig, 1.

Trochammina pseudoinflata
Trochaminna nitida BRADY.- LAGOE 1977, p. 116, pl. 1, figs. 3,4,7.
Trochammina inflata BRADY.- SCHROEDER ET AL 1990, p. 36, pl. 3, figs. 9, 10; pl. 9, figs. 24-26.
Trochammina pseudoinflata SCOTT and VILKS 1991, p. 31, pl. 2, figs. 3-6.

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APPENDIX A

Data Table

Depth in core (cm)	1-3	10-12	20-22	30-32	40-42	48-50	62-64	70-72	81-83	90-92	100-102	111-113	123-125	140-142	160-162	180-182	200-202	220-222	240-242	260-262	280-282	300-302	320-322
Total number of individuals	4288	4224	9288	6240	8320	11208	12152	14440	14576	6088	4776	1112	135	166	167	72	139	207	156	66	132	137	17
Total number of species	15	15	18	17	22	25	27	26	30	23	21	11	2	2	3	3	2	2	3	2	9	8	1
Astracolus hyalancrulus						0.2	0.1	0.1	0.1	0.1													
Bolivina subaenariensis	0.7	8.3	18.7	11.5	14.7	5.8	2.9	5.2	8.1	2.2	1.3	0.7											
Brizalina pseudopunctata		0.6	0.3	0.9	0.2	0.4	1.4	1.0	1.0	0.4	1.3	2.9									3.8		
Buccella frigida					0.1	0.5	0.1			1.6	1.5	0.4									0.8		
Bulimina marginata	20.1	13.4	7.8	8.8	8.0	18.0	13.8	17.3	18.9	12.2	10.7			1.2									
Cassidulina laevigata	15.5	15.9	16.2	16.0	18.0	15.4	17.3	10.0	8.3	7.1	7.0	9.0											
C. reniforme																					2.3	3.7	
Chilostomella oolina				0.4	0.4	0.6	0.4	0.7	1.2	1.1	0.3										1.5		
Cibicides pseudoungerianus			0.1	0.1	0.5	2.6	2.4	1.0	1.1	0.3	0.2												
Cribrostomoides jeffreysi	0.2					0.1																	
Cyclogyra involvens							0.2		0.4	1.1	0.3												
Dentalina spp.								0.2	0.2														
Eggerella advena																							
Elphidium excavatum f. clavatum	9.3	12.9	9.5	13.3	10.9	7.7	7.6	4.4	5.4	7.5	11.4	20.1	97.0	98.8	95.2	94.4	99.3	98.1	97.4	93.9	12.1	65.7	100.0
Fissurina marginata						0.1	0.1	0.2															
Fursenkoina fusiformis	11.0	12.3	11.5	11.3	15.7	24.1	25.1	40.7	23.5	33.8	25.0	36.7									34.1	2.2	
Globobulimina auriculata	9.3	7.6	5.3	5.1	1.8	1.4	2.1	0.7	2.4	4.2	2.7	2.2											
Gyroidina neosoldanii					0.1	0.3	0.1	0.3	0.2	0.3	0.2								0.6			1.5	
Hoeglundina elegans			0.3		0.1	1.3	2.3	0.9	2.6	0.7	0.3				3.0	2.8						1.5	
Islandiella teretis	13.6	8.9	5.9	8.2	9.7	5.1	9.5	6.6	10.6	9.3	19.8	22.3									38.6	22.6	
Lagena spp.	0.4	0.9	0.9	1.5	0.8	0.8	0.5	0.3	0.5	0.1	0.3												
Lenticulina peregrina							0.1	0.2	0.6	1.1	3.4	2.5									1.5		
Nonionella turgida	0.4	2.8	4.7	5.9	5.1	5.3	4.3	4.8	6.4	8.3	8.0												
Nonionellina labradorica	12.1	11.4	14.0	11.4	8.7	2.5	3.6	2.6	2.5	3.2	2.5	1.4									5.3	1.5	
Pullenia subcarinata			1.2	0.4	1.7	1.0	0.6	1.8	2.9	3.4	3.0												
Quinqueloculina agglutinata					0.1				0.1														
Q. seminulum		0.2			0.1	0.7	1.3	0.8	0.4	1.6													
Recurvoides turbinates	0.2								0.1														
Reophax bacillaris						1.5	0.7	0.3	0.4														
R. scorpiurus							0.2		0.1														
R. scotti	0.2	0.6	0.4		0.9		0.1		0.1			1.8	3.0		1.8	2.8	0.7	1.9	1.9	6.1		1.5	
Saccammina difflugiformis	4.3	3.8	2.8	4.0	2.4	2.1	1.7	0.6	1.3	0.4													
Siphotextularia rolshauseni											0.2												
Spiroplectammina biformis																							
Trifarina angulosa			0.2	0.1	0.2	2.1	1.6	0.4	0.4	0.3	0.5												
Trochammina pseudoinflata	2.6	0.4	0.2	0.9		0.4		0.9	0.3														
Total number of planktonics	136	408	728	512	896	2736	1712	1824	1592	1000	384	52											
Total number of ostracods	16	56	56	32	40	56	48	200	616	400	208										3		

Depth in core (cm)	330-332	340-342	350-352	360-362	372-374	383-385	390-392	400-402	410-412	420-422	430-432	440-442	450-452	460-462	470-472	480-482	490-492	500-502	510-512	520-522	530-532	540-542	560-562	570-572
Total number of individuals	126	305	696	316	6396	2576	4936	2248	2240	1984	3768	2248	424	736	2628	805	1660	113	2592	2848	155	332	116	179
Total number of species	3	5	14	3	12	11	14	12	11	11	11	12	5	7	12	12	11	3	11	14	3	7	8	6
Astracolus hyalancrulus																								
Bolivina subaenariensis					0.3				0.4							0.7								
Brizalina pseudopunctata			2.5		0.8	2.8	2.8	4.3	2.9	4.0	1.9	3.2			3.4	0.7	3.1		2.3	2.5				
Buccella frigida			0.1		0.1	1.9	0.5			1.6									0.2				1.7	
Bulimina marginata					0.1											0.3								
Cassidulina laevigata					7.8															0.1				1.7
C. reniforme		2.0	9.6			11.8	15.9	11.4	10.7	13.7	35.0	7.1	4.4	12.5	22.7	12.3	12.5		20.1	7.9		22.3	6.0	3.9
Chilostomella oolina			0.1							0.8	0.4	0.4			0.2		0.7		2.3	2.9				
Cibicides pseudoungerianus																								
Cribrostomoides jeffreysi																								
Cyclogyra involvens																								
Dentalina spp.																								
Eggerella advena	0.8		0.1																					
Elphidium excavatum f. clavatum	95.2	83.3	53.0	94.3	3.7	23.0	23.7	39.5	35.4	36.3	27.0	31.7	70.4	81.5	34.6	63.9	14.2	81.4	23.0	4.8	96.8	70.8	42.2	68.7
Fissurina marginata																								
Fursenkoina fusiformis			13.0	1.3	57.4	37.0	24.3	5.3	4.3	7.3	15.7	23.8	3.1	1.6	27.3	3.0	25.5	6.2	7.7	39.0	2.0	2.1	9.5	6.7
Globobulimina auriculata					0.3	0.9	0.5		1.1	1.2		0.4			0.3					0.1				
Gyroidina neosoldanii			1.2		0.3		1.0	1.8	1.4	3.6	0.6	1.8		0.5	1.1	1.7	0.7		0.5	0.6		0.3	0.9	
Hoeglundina elegans								0.4	0.4		0.9					0.3								
Islandiella teretis		9.2	14.8		23.1	7.1	7.9	12.5	27.1	16.1	13.6	15.3	10.1	1.1	6.2	8.6	37.1		40.0	38.5		1.5	28.5	15.1
Lagena spp.					0.3	2.5	0.8	0.7			0.4	0.7			0.3	0.3	1.5		1.2	0.7			1.7	
Lenticulina peregrina						1.9	2.6	2.1	2.5	3.6	1.5	1.4			0.5	0.7	0.5		0.6	2.2				
Nonionella turgida																								
Nonionellina labradorica		3.0	3.2		6.1	10.3	18.8	21.4	13.9	11.7	3.0	13.5	11.9	1.6	3.5	7.6	3.4		1.9	0.3	1.3			
Pullenia subcarinata																								
Quinqueloculina agglutinata																								
Q. seminulum						0.9	0.7	0.4				0.7			0.2					0.1		0.3		
Recurvoides turbinates																								
Reophax bacillaris																								
R. scorpiurus																								
R. scotti	4.0	2.6	1.4	4.4			0.3							1.1			0.7	12.4				2.7	9.5	3.9
Saccammina difflugiformis																								
Siphotextularia rolshauseni																								
Spiroplectammina biformis																								
Trifarina angulosa			0.1				0.3	0.4												0.1				
Trochammina pseudoinflata																								
Total number of planktonics							8	32	48	24		8	11			3			12	8				51
Total number of ostracods			1		24	56	40				8				8	5	4							

Depth in core (cm)	580-582	590-592	600-602	610-612	620-622	630-632	640-642	650-652	660-662	670-672	680-682	690-692	700-702	710-712	720-722	730-732	740-742	750-752	760-762	770-772	780-782	790-792	800-802
Total number of individuals	103	162	594	3168	1122	773	220	1758	2856	2682	2316	2670	2040	2256	2274	1800	2952	2808	2514	3828	3336	5310	1506
Total number of species	2	4	10	6	6	5	2	7	9	7	9	7	6	9	5	4	6	8	5	6	7	7	7
Astracolus hyalancrulus																							
Bolivina subaenariensis																							
Brizalina pseudopunctata			0.2	2.8				0.3	2.0	0.2	0.1			0.3			0.2						
Buccella frigida			1.0					0.3	0.6														
Bulimina marginata					0.3																		
Cassidulina laevigata			0.8	1.3																			
C. reniforme		3.7	3.0	12.1	1.3	1.4		10.9	26.3	47.4	32.6	14.6	21.2	21.3	20.6	22.0	16.7	17.7	11.2	15.2	23.6	12.1	26.7
Chilostomella oolina									0.3		0.1							0.4					
Cibicides pseudoungerianus																							
Cribrostomoides jeffreysi																							
Cyclogyra involvens																							
Dentalina spp.																							
Eggerella advena					0.8	6.4	15.5	10.6	4.2	1.8	1.2	1.8	1.2	1.6	1.8		0.4	1.1		1.9	1.4	0.8	2.4
Elphidium excavatum f. clavatum	98.1	90.7	61.5	77.3	94.7	91.3	84.6	77.1	63.9	48.3	61.3	82.5	74.7	74.5	76.8	77.0	82.1	79.5	87.8	81.7	72.7	85.4	66.1
Fissurina marginata																							
Fursenkoina fusiformis		1.9	1.7	1.8					0.6	0.5	2.5	0.4	1.5	1.3				0.4			0.7		2.0
Globobulimina auriculata			0.2																				
Gyroidina neosoldanii											0.1											0.1	
Hoeglundina elegans																							
Islandiella teretis			26.3	4.8	2.4	0.8		0.3	2.0	1.3	1.9	0.2	1.2			0.8	0.2	0.4	0.2	0.5	0.5	1.4	1.6
Lagena spp.									0.3	0.5	0.1	0.2	0.3	0.3	0.3		0.4	0.2	0.5	0.6	0.5		0.4
Lenticulina peregrina			0.8																		0.5	0.1	
Nonionella turgida																							
Nonionellina labradorica																							
Pullenia subcarinata																							
Quinqueloculina agglutinata																							
Q. seminulum																0.2							
Recurvoides turbinates																							
Reophax bacillaris																							
R. scorpiurus																							
R. scotti	1.9	3.7	4.6		0.8	0.2		0.3															
Saccammina difflugiformis																							
Siphotextularia rolshauseni												0.2		0.3	0.5			0.2		0.2		0.1	0.8
Spiroplectammina biformis														0.3									
Trifarina angulosa																							
Trochammina pseudoinflata														0.3									
Total number of planktonics												12			6	12	6	6					
Total number of ostracods			4									18		6			6	12				6	6

APPENDIX B

AMS ^{14}C Dates

Depth in core (cm)	2.4	99	295
Source	mollusc shell	foraminifera	mollusc shell
AMS C-14 Age	3970	9220	11340
Error (+/-years)	60	90	70
Lab Number	Beta-61398	Beta-61624	Beta-57906
Depth in core (cm)	345	345	416
Source	mollusc shell	mollusc shell	foraminifera
AMS C-14 Age	10740	10870	12420
Error (+/-years)	90	70	80
Lab Number	Beta-57905	Beta-57907	Beta-61626
Depth in core (cm)	590	616	695
Source	mollusc shell	mollusc shell	mollusc shell
AMS C-14 Age	12020	12230	13050
Error (+/-years)	90	70	140
Lab Number	Beta-57908	Beta-57909	GX-17635