CONTROLS ON WESTPHALIAN PEAT ACCUMULATION:

THE SPRINGHILL COALFIELD, NOVA SCOTIA

by

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For Jane, and my Parents

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ABSTRACT

During the early Westphalian (Early to Middle Pennsylvanian), tropical peatlands (mires) flourished along the southern margin of the intermontane Cumberland Basin in the area of the present day Springhill coalfield. The resulting bituminous coal seams and associated sideritic mudrocks and multistorey channel sandstone bodies (Springhill Mines Formation) lie basinward of, and progressively onlap, alluvial fan conglomerates (Polly Brook Formation) derived from the adjacent Cobequid Highlands. Within the basin-fill sequence, coals occur in association with grey mudrock and multistorey sandstones (cyclothems, 20-30 m thick). These strata are constrained within a 600 m-thick "coal window" within the fining and reddening upward basin-fill sequence (2-5 km thick). The Springhill coals, up to 4.3 m thick and spanning a transverse distance of 5 to 8 km, exhibit a distinct areal zonation. The piedmont zone is marked by ill sorted partings attributed to distal alluvial fan sheetfloods. The inner mire zone contains few inorganic partings and is lowest in ash and sulphur. The riverine zone is characterized by splits which locally entomb sandstone bodies of basin-axis rivers.

The precursors to the Springhill coals were rheotrophic (flow fed) peatlands, fed by nutrient-rich groundwater emanating from the fans. Evidence includes the consistent relation between coals and basin-margin conglomerates, the incursion into the mires of distal fan sheetfloods, trends in lithotype and maceral composition of the coals, especially gelification of vitrinite macerals, and the domination of mire flora by arboreous lycopsids suited to areas of nutrient-rich groundwater influx. The rheotrophic mires, and their reliance on fan discharge, indirectly suggest a sub-humid tropical paleoclimate.

A number of controls interacted to influence the development of the peat-forming ecosystems. Tectonism exerted a major control at the basin-fill sequence level (10^6 yr.) through both extremely rapid subsidence and orogenesis. The resulting alluvial fans discharged nutrient-rich groundwater supply which in modern sub-humid settings is crucial to the maintenance of wetlands in drier periods. The *degree* to which tectonism influenced mire development through groundwater discharge was therefore linked inextricably to the prevailing climate. Progressive denudation of the Cobequid Highlands probably resulted in a decreased catchment area, orographic rainshadow and fan size and may have contributed to the progressive decline in peat accumulation. The coal window within the basin-fill sequence is inferred to reflect a coincidence of optimal groundwater supply and net subsidence. Cyclothems indicate that conditions favouring peat accumulation recurred with a cyclicity in the order of 10^4yr. The similar timeframe of Milankovitch cycles of obliquity and precession suggest that climatic or related (eustasy, geoid) change was influential in cyclothem and mire development.

ACKNOWLEDGEMENTS

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CHAPTER 1: INTRODUCTION

1.0 THE QUESTION

Since beginning work in the assessment of coal resources in the Springhill coalfield in 1977, the author has struggled with the vexing question of their origin. Apart from being a stimulating academic quest, knowledge of the controls on location, form and chemical character of the coals is a prerequisite to formulation of sound exploration models. The research summarized in this dissertation stems from a dissatisfaction with many published "models of coal deposition" which seemed to the author to be either one dimensional or frustratingly vague.

A basic premise of this research is that paleoenvironmental controls will have had a hierarchial impact on ancient mire development. The approach taken in attempting to unravel this hierarchy centers around the concept of coal as the product of a peat-forming wetland ecosystem. The term **ecosystem**, introduced by Tansley (1935, p. 299), refers to "the whole *system* (in the sense of physics), including not only the organism-complex, but also the whole complex of physical factors forming what we call the environment of the biome - the habitat factors in the widest sense." The acceptance of the ecosystem concept virtually assures a multiplicity of controls, as these ancient peatlands (mires) interacted with all aspects of their environment. A systematic examination was therefore undertaken of geological, biological and climatic influences on the development of these ancient peatlands.

A wide range of controls on ancient peat formation have been identified or proposed in the literature. By far the majority invoke *allogenic* (external) controls, such as: tectonism/subsidence (Shaw, 1951; Haites, 1951; Copeland, 1959; McLean and Jerzykiewicz, 1978; Bless *et al.*, 1981; Weisenfluh and Ferm, 1984; Li Sitian *et al.*, 1984; Courel *et al.*, 1986; Wood *et al.*, 1986; Courel, 1988; Vallé *et al.*, 1988; Naylor *et al.*, 1989; Yeo, 1988; McCabe, 1991, among others); glacioeustasy, commonly attributed by later authors to astronomically-driven climatic cycles (Wanless and Shepard, 1936; Fischer, 1986; Heckel, 1986; Cross, 1988; Ross and Ross, 1988, among others); climate (Smith, 1962; Cecil *et al.*, 1985; Zeigler *et al.*, 1987, Zeigler, 1990; Guerra-Somer, 1991); autocyclic sedimentary mechanisms (Kosters *et al.*, 1987; Flores, 1981; Rust *et al.*, 1987); basin hydrology (Jackson, 1979; Ethridge *et al.*, 1981; Galloway and Hobday, 1983); and differential compaction (Fielding, 1984; Courel, 1987; Moore *et al.*, 1990). Some of these workers have identified a combination of controls. *Autogenic* (internal) controls have been proposed as well; perhaps the most commonly cited of these is the protection from clastic deposition provided by doming of raised (ombrogenous) mires (Flores, 1980, 1981; Smyth, 1980; McCabe, 1984; Warwick and Flores, 1987).

There are at least two reasons apparent for the diversity of proposed controls. One is that the most obvious controls may actually vary with paleogeographic setting (e.g. see Hacquebard *et al.*, 1967; Klein and Willard, 1989). A second is that although a multitude of controls existed for a given area of ancient peat accumulation (e.g. Teichmüller, 1962; Calder, 1986; Hobday, 1987; Guion and Fielding, 1988; Flores, 1989), there have been few attempts at methodical assessment of the entire hierarchy from basin to intra-seam scale. Quite possibly the required integration of scientific disciplines has acted as a deterrent. The control identified in some cases may correlate closely with the investigator's field of expertise.

Various published concepts have shaped the research described in this dissertation. A long-known but important point is the recognition of a multitude of controls on coal formation (Teichmüller, 1962, 1982) as noted above. These controls can be ranked in hierarchial order (Calder, 1986; Guion and Fielding, 1988) within a temporal framework (Heckel, 1986; Leeder and McMahon, 1988; Chesnut, 1989; Klein, 1990). Importantly, basinal and peat-forming processes may be expected to interact (Flores, 1981; Galloway and Hobday, 1983; Warwick and Flores, 1987). An important concept is that peat formation may not always be allocyclically terminated but may be self-limiting (Clymo, 1987), a theory supported by studies of modern tropical forest peatlands, which show ever-decreasing rates of peat accumulation in response to decreasing fertility as a consequence of peat doming (Anderson and Muller, 1975; Anderson, 1983).

As geologists, we are at least in part bound to the Law of Actualism, therefore studies of modern peat-forming systems as analogues for coal formation are particularly important (Spackman, 1958; Spackman *et al.*, 1969; Cohen, 1974; Cohen and Spackman, 1972; Cohen and Spackman, 1980; Cohen *et al.*, 1987; Moore, 1987; Cameron *et al.*, 1989; Esterle et al., 1989; Moore and Hilbert, in press). Building on these studies, recent research has focused on deduction of ancient peat-forming processes (Smith, 1962; Esterle and Ferm, 1986; Bartram, 1987; Fulton, 1987; Warwick and Stanton, 1988; Esterle *et al.*, 1989; Eble and Grady, 1989; Moore *et al*, 1990). While not always in agreement, the research summarized in this dissertation has been shaped by discussions and interaction with many of these authors. Processes of ancient peatland development must be integrated, however, with studies of the physiology, function and ecology of the Carboniferous peatland floras (Scott, 1978, 1979; DiMichele and Phillips, 1985; DiMichele and Demaris, 1987; DiMichele and Phillips, 1988; DiMichele *et al.*, 1985; Collinson and Scott, 1987; Phillips, 1979; Phillips *et al.*, 1985, among others).

The author's research has drawn heavily upon studies of modern peat-forming systems, in particular the treatise edited by Gore (1983), for the development of methods by which ancient processes and controls can be tested. The current focus on destruction of modern tropical rainforest ecosystems provides some hint of both the intricacy of nature's balance and the limits of our knowledge even of possible modern analogues. The degree to which our understanding of controls on ancient peat accumulation can be furthered is limited by the extent of our knowledge of local controlling elements such as basin tectonics and paleoclimate. Nonetheless, it is the hope of the author that the seeming Gordian knot of controls on coal formation, if not unravelled, will at least be loosened by this research.

1.1 CONTENTS OF THE DISSERTATION

This introductory chapter describes the method of approach, including sources of data, status of geological investigations prior to this study and geological setting of the Springhill coalfield. The sedimentology of the coal-bearing basin-fill is described in Chapter II in the context of lithofacies assemblages. These fundamental components of paleogeographic reconstruction are placed in their spatial (stratigraphic) context in Chapter III, enabling reconstruction of the paleogeographic setting of the peatlands, and identification of controls on their distribution evident from basin-fill analysis. The deposit of one of these peat-forming ecosystems, the No. 3 seam, is examined in detail in the subsequent chapter (IV), from which the processes of mire development and environmental interaction are deduced. In the concluding chapter (V), the various controls are discussed within a temporal framework. Paleoclimate is reconsidered in this concluding chapter, in the light of considerations discussed in the previous chapters.

1.2 PEATLANDS: TERMS AND DEFINITIONS

Ecosystem	a biological community and the physical environment with which it interacts
<u>Wetlands: general terms</u>	
mire	a general term used to describe any wetland ecosystem that is peat-forming (Moore, 1987)
fen	a flow-fed (rheotrophic) mire for which the surface is above the dry-season water table (Tansley, 1939). Note, however, that the term fen was used by Moore and Bellamy (1974) and Gore (1983) to

	denote all rheotrophic mires, including swamps (see Table 4.1).
swamp	a flow-fed (rheotrophic) mire for which the surface is perennially below the water table, even during dry season (Tansley, 1911).
bog	a mire that is wholly rain-fed (ombrotrophic).

<u>Nutrient Supply:</u>

ombrotrophic	a wetland ecosystem, usually raised or domed, that receives its nutrient supply solely from precipitation.
rheotrophic	a wetland ecosystem, usually low-lying (cf. planar), that receives nutrients from groundwater (flow-fed). Note: Rheotrophic ecosystems receive recharge both from precipitation and groundwater.
mesotrophic	an intermediate classification between ombrotrophic and rheotrophic; a wetland tending toward more ombrotrophic conditions but still receiving some nutrient input from groundwater.
eutrophic	mineral- and nutrient-rich.

<u>Processes:</u>

terrestrialization	a self-induced (autogenic) mechanism of mire development by which the peat surface rises relative to the water table (the German Verlandung of Weber, 1908).
paludification	a mechanism of mire development by which wet conditions are induced at a site, rendering it favourable for mire expansion (the German Versumpfung of Weber, 1908).
autogenic	self-induced, internal (change, etc.).
allogenic	produced by external influences.

<u>Plant Communities:</u>

limnetic, limnic	a plant community in which the flora are free-floating or rooted below water.
telmatic	a plant community rooted at the water table.
terrestric	a plant community rooted above the maximum level of seasonal flooding.
semi-terrestric	a plant community rooted above the water-table but below the

	maximum level of flooding.

Topographic and Sectional Terms:

topogenous	a mire that is low-lying (cf. planar)
ombrogenous	a mire having a surface raised above groundwater level (cf. domed, or raised bog).
rand	the outward-sloping perimeter of a raised mire.
acrotelm	the surficial, active layer of peat which is relatively more water-permeable (Ingram, 1978).
catotelm	the sub-surface, "fossilized" layers of peat (Ingram, 1978).

1.3 SPRINGHILL COALFIELD: LOCATION AND TOPOGRAPHIC SETTING

The Springhill coalfield is situated in the south-center of the Cumberland Basin, northern Nova Scotia, (Figure 1.1). The known area of the coalfield is approximately 65 km^2. Much of the coalfield and seam subcrops underlie the town of Springhill (Figure 1.2). Springhill is built on the summit and western flank of a hill which reaches a maximum elevation of 198 m above sea level. The topography is generally characterized by gently undulating hills, however to the east the land falls away abruptly to elevations near 70 m in the Black River Valley, a nearly linear southwest-northeast trending feature. At a point immediately south of the Black River Road, the drainage in the valley divides, with headwaters of the Black River flowing northeasterly through Salt Springs, and headwaters of the Southampton River flowing southwesterly toward Leamington. Fluvio-glacial sand and gravel deposits, approximately 10 m thick, characterize the valley north of the drainage divide. The hilly environs south of the town near the community of Rodney are a combination of pasture, blueberry fields, and mixed deciduous-coniferous woodland. Further south, the topography becomes significantly higher in elevation as the more resistant rocks of the Cobequid highlands are encountered.

The western side of Springhill is drained by the headwaters of a series of westerly-flowing streams, including (from north to south) East Branch Southampton River, Mills Brook, Coal Mine Brook, Harrison (or Mountain) Brook, and Maccan River. Drainage is poor in the eastern environs due in part to a 0.5 m thick clay-rich soil layer that overlies approximately 3 - 4 m of glacial till. The land here is characterized by open fields, eventually giving way to mixed coniferous-deciduous woodland to the east. In the extreme north of the coalfield near highway No. 2, dissolution of Early Carboniferous evaporites has resulted in the formation of sinkhole lakes. Outcrop is generally restricted to stream beds, with Black River, Polly Brook and Harrison Brook providing the best

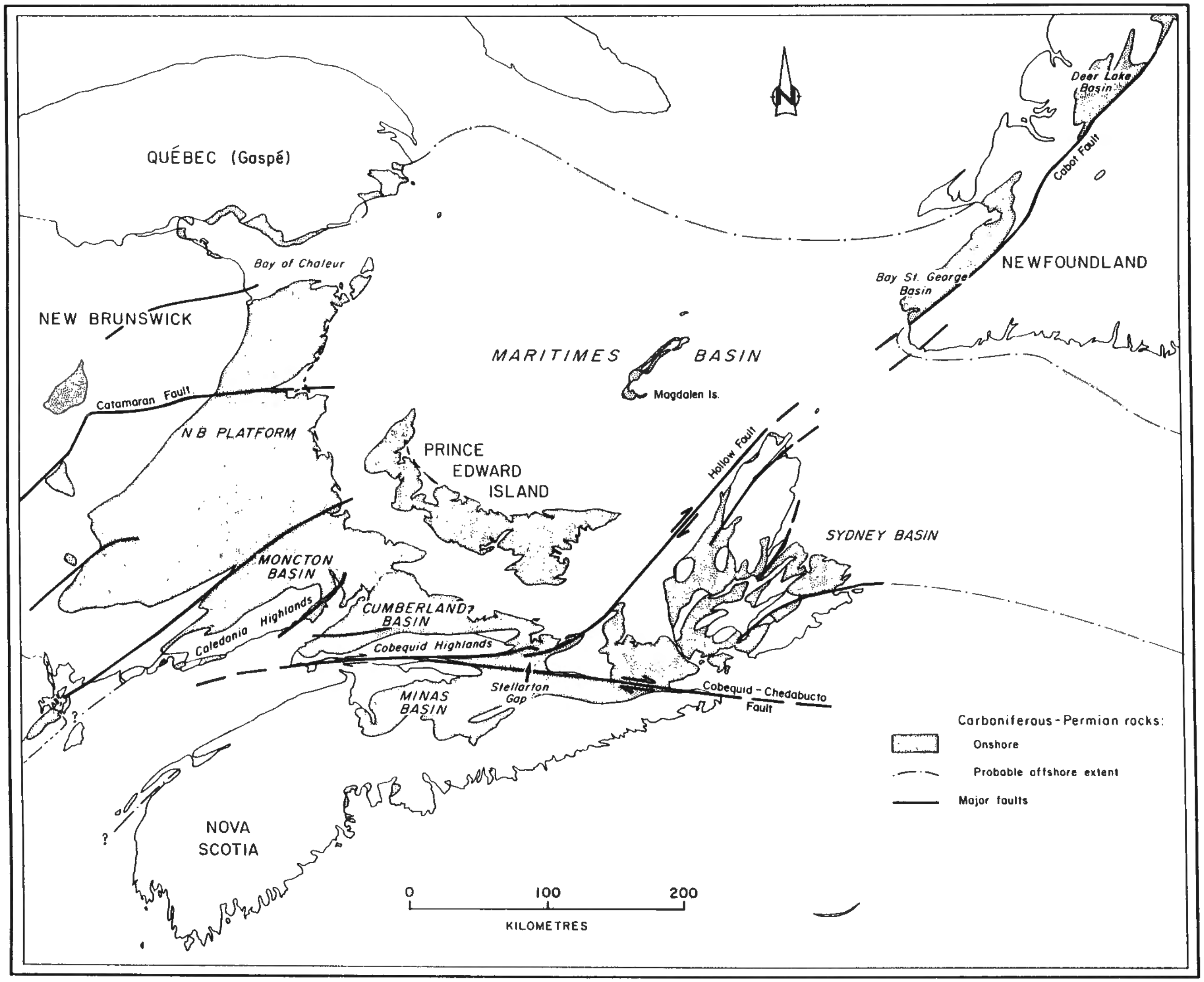


Figure 1.1 The Maritimes Basin of eastern Canada

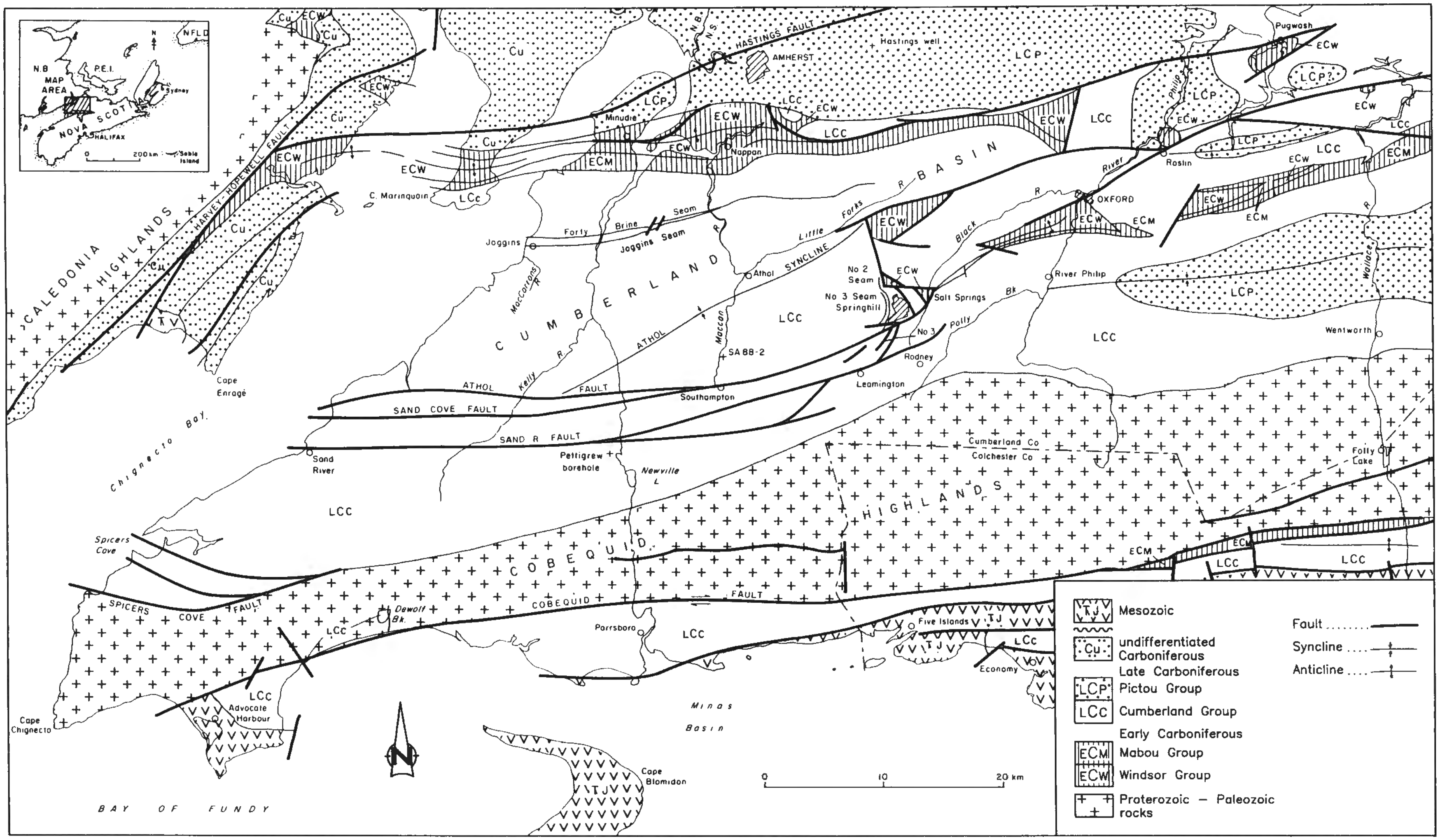


Figure 1.2 Geology of the western Cumberland Basin (after Calder, 1984c; Ryan *et al.*, 1990).

exposures. Access to low-lying exposures on Coal Mine Brook, described by Fletcher (in Kerr, 1924), is hindered by sewage contamination.

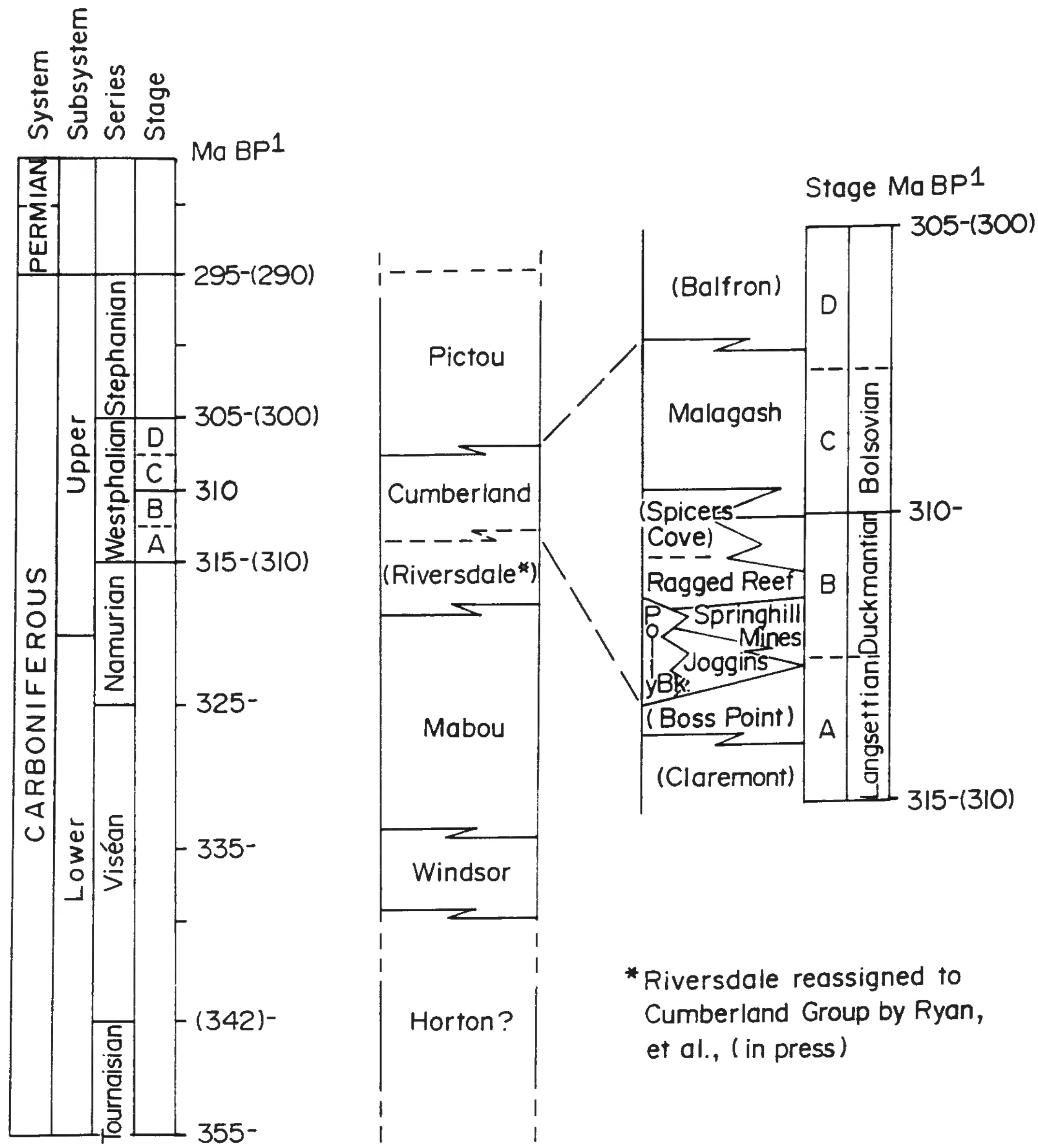
1.4 GEOLOGICAL SETTING OF THE CUMBERLAND BASIN

The Springhill coalfield occurs in the south of the Cumberland Basin. The Cumberland Basin is a component of the much larger and variously disrupted Maritimes Basin of eastern Canada (Figure 1.1). Carboniferous strata dominate the basin-fill which ranges from Devonian to Permian (Figure 1.3).

The extent of the Cumberland Basin has not been accurately defined. To the west and south, the basin is bounded by the Caledonia and Cobequid highland massifs, but to the north and east, the limits of the basin are obscured by the broad regional cover of younger (Westphalian C-Permian; Barss *et al.*, 1963; Barss and Hacquebard, 1967) strata assigned to the Pictou Group and by waters of the Gulf of St. Lawrence. The known areal extent of the basin is 5000 km^2.

The Cumberland Basin comprises several east- to northeast-trending synclines separated by salt anticlines cored with early Carboniferous marine evaporites of the Windsor Group. The basin has thus been referred to as the Cumberland Synclinorium (Keppie, 1982a), although this name has not gained wide usage. The salt anticlines are thought to have formed adjacent to and over basement horsts which suggests that the component synclines in fact may have been individual sub-basins. This is especially true for the Athol Syncline (Figure 1.2) which is host to the Springhill and Joggins-Chignecto coalfields. The great thickness of Westphalian B strata within the syncline and thickening of stratigraphic units coincidentally with the synclinal axis evident in high resolution seismic profiles (Bromley and Calder, in press) are evidence that the Athol Syncline was a local depocenter and indeed a basin in its own right. The Athol Syncline, a 25 x 70 km rhomb, shares the western and southern boundaries of the Cumberland Basin. To the north, the syncline shares a common limb of the Minudie Anticline and to the east is bounded by faults associated with the salt wall (Black River diapir) terminus of the Claremont Anticline.

The south limb of the syncline is truncated by a recently discovered major east to northeast-trending structure, the Athol-Sand Cove fault zone (Bromley and Calder, in press; Ryan *et al.*, 1990). Northeast trending faults which disrupt the coal measures of Springhill are believed to be components of the Athol-Sand Cove fault zone (Calder and Bromley, 1988).



1 - Ar^{40}/Ar^{39} dates for Upper Carboniferous from Lippolt and Hess (1985); alternate dates ( ) from IUGS (1989).

Figure 1.3 Stratigraphic column, Cumberland Basin, with formations of the Cumberland Group.

1.5 REGIONAL TECTONIC SETTING

The Maritimes Basin (Roliff, 1962; Williams, 1974; Boehner *et al.*, 1988) of which the Cumberland Basin is a component (Figure 1.1), is thought to have been formed in an intracontinental transform zone, characterized by strike-slip transpressional deformation and rapidly subsiding extensional sub-basins (Bradley, 1982). It originated near the end of the Acadian Orogeny in the mid to late Devonian (Poole, 1967). The Acadian orogeny witnessed the accretion of the Avalon and Meguma terranes, part of Gondwana (Schenk, 1971, 1981; Johnson and van der Voo, 1986), to the North American craton, Laurussia. The Maritimes Basin occurs within a major strike-slip zone which includes the Variscan Belt of Europe and North Africa. During the late Paleozoic, this predominantly transcurrent zone connected the compressional belts of the Appalachians and Urals (Arthaud and Matte, 1977; Heward and Reading, 1980). The basin has been attributed to various origins including: rift (Belt, 1968), strike-slip pull-apart (Bradley, 1982), passive subsidence (McCutcheon and Robinson, 1987) and thrust sheet re-extension (Quinlan, 1988).

During the Hercynian-Alleghenian (cf. Variscan) orogeny dextral strike-slip along major east-west faults deformed the southern part of the Maritimes Basin (Eisbacher, 1969; Keppie, 1982b; Donohoe and Wallace, 1985; Nance, 1987). The major fault system of the southern Maritimes Basin, the Cobequid-Chedabucto (Minas Geofracture of Keppie, 1982b), played an important role in the evolution of adjoining subbasins (Figure 1.4). The Cobequid fault was part of a major décollement along which the Meguma Terrane moved and rotated eastward, with up to 150-200 km of dextral offset through the Late Carboniferous and Permian (Keppie, 1982b; Scotese *et al.*, 1984).

This dextral displacement along the Cobequid Fault is thought to have been partially accommodated by a change from a near-vertical wrench along the Cobequid highlands south of the Cumberland Basin (Webb, 1969; Donohoe and Wallace, 1985) to a listric thrust further west near southern New Brunswick, at the southwestern margin of the Cumberland Basin (Nance, 1987). This transpressional thrusting resulted in alluvial fan building in the Tynemouth Creek Formation of Westphalian age (Plint and van de Poll, 1984; Dolby, 1988b). Further east, near the terminus of the Cobequid fault, dextral strike-slip of the Cobequid and Hollow wrench faults led to the pull-apart formation and syntectonic infilling of the Stellarton Basin/Pictou coalfield (Fralick and Schenk, 1981; Yeo and Ruixiang, 1987) predominantly during the Westphalian C (Dolby, 1986a). Hence, the Cobequid Fault was active throughout the time period of deposition at Springhill and Joggins (Westphalian A-B).

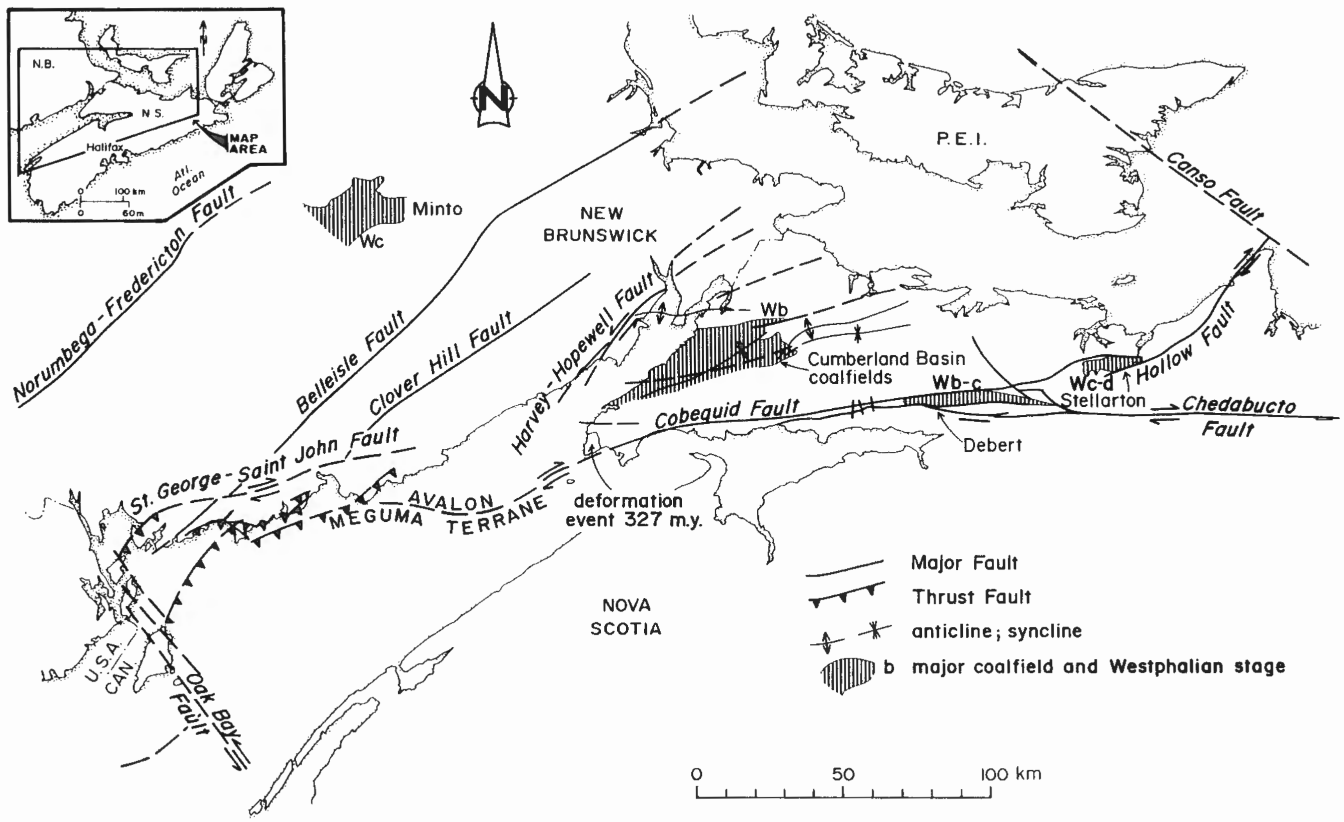


Figure 1.4 Tectonic setting of coalfields, southwestern Maritimes Basin. (modified from McCutcheon and Robinson, 1987; Yeo and Ruixiang, 1987).

The development of thick Westphalian A-B basin-fills adjacent to the Cobequid, Athol-Sand Cove, Hollow faults in northern Nova Scotia (Figure 1.1 and 1.4) and the Cabot fault in western Newfoundland (Figure 1.1) suggests that this was a major tectonic episode during development of the Maritimes Basin. There is a general scarcity throughout the Maritimes Basin, of Westphalian B sedimentary cover especially, except in these localized depocenters. This further suggests the early Westphalian was a time of regional erosion. If the Athol-Sand Cove fault zone proves to have been active during this time, it would have had a profound effect on the regional paleogeography, basin-fill and peat-forming ecosystems.

Tectonism is a potential major influence on development of the ancient peatland systems of the Springhill coalfield. Unfortunately, the origin of the Cumberland Basin from a tectonic standpoint is poorly understood beyond its occurrence in a region dominated by strike-slip faulting. The Cumberland Basin fill, however, does meet most, if not all, criteria for a strike-slip regime as given by Mitchell and Reading (1986): i) extreme lateral facies changes; ii) very great thickness of rapidly deposited sediment; iii) abundant sediment supply from multiple sources; iv) unconformities; and possibly, v) contemporaneous deformation.

The eastern Athol Syncline, including the Springhill coalfield, also exhibits characteristics of an extensional setting. Experimental modelling of listric extensional fault systems (Ellis and McClay, 1988) has revealed three major structural elements: i) a crestal collapse anticline; ii) a non-rotational translational block; and iii) a zone of rotation nearest the detachment breakway. The Springhill coalfield is dominated by such a crestal collapse anticline (Calder, 1984a). High resolution reflection seismic profiles north of the anticline suggest a translational element to the basin-fill (Bromley and Calder, in press). To the east of the coalfield, pronounced angular unconformity between Riversdale and Cumberland Group strata on Black River (Bell, 1938) suggest a roll-over component with a detachment adjacent to the Cobequid highlands, but the fact that the Riversdale and Cumberland strata are in fault contact as mapped by the author casts doubt on such an interpretation. In addition, the role of salt kinetics in the development of the structural elements must be considered.

Basin asymmetry suggested by the abundance of coarse margin deposits at Springhill and the virtual absence of such deposits, coupled with southeasterly paleoflow at Joggins toward the synclinal (basin?) axis, has been recently determined to be characteristic of rifts (Rosendhal *et al.*, 1986; Frostick and Reid, 1987). The northern limits of the coal basin and the relative position of the Joggins coal measures have yet to be determined. Finally, the initial interpretation of the geometry of the Athol-Sand Cove fault (Calder and Bromley, 1988; Ryan *et al.*, 1990) raises the possibility that the Springhill coalfield developed in a pull-apart setting. Obviously, much remains to be clarified.

1.6 PALEOGEOGRAPHIC AND PALEOCLIMATIC SETTING

The coalfields of the Maritimes Basin formed part of a broad expanse of peatlands known as the Euramerican coal province, which flourished over much of the continent of Laurussia during the Late Carboniferous. The ancient peatlands of the Cumberland Basin occupied a sub-equatorial, tropical position during the Westphalian B (Scotese *et al.*, 1979). By Westphalian C-D time, northward movement of Laurussia had placed the basin near the 5° S paleolatitude (Scotese et al., 1979; Rowley et al., 1985).

The coal-bearing strata of the Cumberland Basin have yielded no macroscopic evidence of marine conditions, and the position of the ocean(s)/sea(s) relative to the basin is largely unknown. Gibling *et al.* (in press) and Gibling and Bird (in press) postulated that open ocean may have existed to the east of the Sydney coalfield during the Westphalian D. Agglutinated foraminifera extracted from coal-bearing strata of the Sydney Mines Formation suggest freshwater to brackish conditions (Thibaudeau and Medioli, 1986; Gibling and Bird, in press). The Appalachian sea is believed to have been southwest of the central Appalachian Basin during the early Westphalian/Pennsylvanian (Donaldson and Shumaker, 1981).

To the southwest of the Maritimes Basin, the rising Appalachian mountains may have been the source of many of the predominantly northeasterly flowing river systems that traversed the basin during the Westphalian (Gibling *et al.*, in press). The Appalachian mountains, which may have attained central Andean dimensions (Slingerland and Furlong, 1989) would have almost certainly affected regional weather patterns as well: the intense low pressure cell that would have developed over the mountains may have contributed to a monsoonal paleoclimate (Rowley *et al.*, 1985).

Studies of Euramerican paleoclimate have focused on the Appalachian Basin; no reconstruction has specifically addressed the paleoclimate of the Maritimes Basin. Interpretations of Westphalian B climate (Figure 1.5) span the spectrum from dry (Phillips and Peppers, 1984) and relatively dry (Winston and Stanton, 1989) but seasonal (Phillips *et al.*, 1985) to everwet tropical (Cecil *et al.*, 1985) with a possible periodic shift to long wet/short dry seasonal (Cecil, 1990). Because much of the research presented here considers evidence of climatic change, the subject of paleoclimate is considered further in the concluding chapter.

1.7 PAST GEOLOGICAL INVESTIGATIONS

The first written report of coal at Springhill was made by Abraham Gesner (1836):

"In the bottom of a small brook, running through a wild forest of beech and maple, a poor farmer has been digging coal, one of the greatest treasures of the earth." Gesner continued, as if to warn following generations of geologists, **"the rocks are so deeply covered with the rubbish of the surface, that the visitor will return from the wilderness disappointed in the collection of fossils, and fatigued by a journey over an uneven surface."**

Hartley (1869), Gilpin (1898, 1901) and Gray (1917) provided early reports on the occurrence and economic aspects of the Springhill coals. The work of Fletcher (1873-1909) constituted the most involved geological assessment of the coalfield to that time, and his 1903 map the first comprehensive documentation of lithology and outcrops. Kerr (1924) expanded upon the work of Fletcher in his unpublished Geological Survey of Canada report. The configuration of seams painstakingly documented in field maps of Fletcher and Kerr has largely survived the test of time.

The foundation of stratigraphic studies in the Cumberland Basin has been the work of Bell (1914-1966). Bell mapped the Cumberland coal basin (Bell, 1938) and undertook detailed studies of the fossil flora (Bell, 1944; 1966). His work continues to be the foundation for subsequent stratigraphic efforts. The stratigraphy and sedimentology of the Cumberland Basin was the subject of the Ph.D. dissertation of Shaw (1951) who consequently authored GSC Preliminary Map 51-11. The most extensive published report to date on the Springhill coalfield is that of Copeland (1959). His GSC memoir and accompanying plans deal both with the Springhill and Joggins-Chignecto coalfields and provide important data on seam structure and lithology from underground mines now inaccessible.

Hacquebard, Birmingham and Donaldson (1967) conducted microlithotype analyses of polished blocks from the No. 2 and No. 3 seams, which constitutes the only petrographic investigation of Springhill coals prior to the author's work. Hacquebard and Donaldson (1964) assessed the palynology of the coal-bearing districts in the Cumberland Basin for the purpose of stratigraphic correlation and incorporated rank data for the No. 2 seam in their work on coalification in the

Maritimes Basin (Hacquebard and Donaldson, 1970). Hacquebard also compiled an extensive unpublished atlas of resource calculations, structural interpretation and borehole correlation following extensive diamond-drilling programs in 1959-61.

Ward (1979) erected a provisional nomenclature for seams correlated on the south limb of the Springhill Anticline during the course of exploratory drilling for opencast resources by G. Wimpey and Associates in 1977-78. This greatly facilitated subsequent efforts by the author linking coal seam stratigraphy of the northern and southern limbs of the Springhill Anticline.

1.8 INVESTIGATIONS BY THE AUTHOR

Prior to the formal commencement of this research, the author assisted in the Wimpey Laboratories open-cast investigation in 1977, continuing thereafter to supervise coal exploration diamond-drilling by the Nova Scotia Department of Mines and Energy in the Cumberland Basin. Much of this effort was expended in the axial region of the Springhill Anticline following the identification of the previously unrecognized and unmined Rodney seam in 1979 (Calder, 1980). It was during the course of drilling in 1979-1981 that much of the geological foundation was laid for this research. A key element of this foundation was re-interpretation of the stratigraphy, including correlation of coal seams (Calder, 1980; 1981a, 1981b, 1981c; and in Norwest, 1981). Simultaneously and necessarily, the structure of the anticlinal axis, marked by complex, brittle deformation and crestal collapse, was largely resolved (Calder, 1980; Norwest, 1981; Calder, 1984a). The informal stratigraphic units introduced by the author (Calder, 1980), have survived largely unmodified to comprise three of the lithofacies assemblages described in this study. The six lithofacies assemblages of this study in turn have served to define the Springhill Mines Formation and its members, as well as the Polly Brook Formation and Leamington Member (Ryan *et al.*, 1990; in press).

Since 1978, 88 coreholes have been drilled in the Springhill coalfield by the Nova Scotia Department of Mines and Energy under the supervision of the author; a further 6 were drilled in the Salt Springs area northeast of Springhill. The author logged 72 of these, totalling 9337 m. This thesis is based to a large degree upon the wealth of sedimentological and coal petrographic data obtained during the course of this exploration activity. The remaining 22 coreholes were logged by R. Naylor, K. Gillis, G. Somers, S. Chesal and C. MacPherson of the Coal Section, NSDME, and by Norwest Resource Consultants. Key sections of these cores, in particular all those employed in detailed study of the No. 3 seam (this study), were also relogged by the author.

The Novaco open-pit mine, mapped on an on-going basis throughout its mining history from 1982-1983, provided important and otherwise unavailable structural, sedimentological and paleoecological data of the Rodney seam and overlying Rodney Sandstone. During the period 1978-1987, the author mapped in detail the 150 km^2 map area of the Springhill coalfield (Calder, 1990). This stratigraphic and structural mapping was included in the 1:50,000 scale geology map (Amherst, Springhill and Parrsboro sheet) of the Cumberland Basin, co-authored with Ryan *et al.* (1990). High resolution reflection seismic surveys run by the Geological Survey of Canada during the period 1985-86 provided previously unavailable data on stratigraphic architecture, structure and correlation of coal-bearing strata throughout the Athol Syncline (Calder, 1987; Calder and Bromley, 1988; Bromley and Calder, in press).

CHAPTER II: SEDIMENTOLOGY OF THE SPRINGHILL COALFIELD: LITHOFACIES ASSEMBLAGES

2.0 INTRODUCTION

This chapter describes the sedimentology of the coal-bearing and coarse basin-margin strata in the area of the Springhill coalfield (NSDME map OFM 90-06). The chapter focuses upon the description and interpretation of seven lithofacies assemblages defined in this study. These assemblages, in the order of 200-500 m thick, are the basis for subsequent paleogeographic interpretation. The lithofacies assemblages are defined on relative proportions of lithofacies, based on data obtained from outcrop mapping and diamond-drill cores (Nova Scotia Department of Mines and Energy Open File Report).

Several of the lithofacies assemblages of this study have been adopted as or used in the definition of lithostratigraphic units of the Cumberland Group (Ryan *et al.*, in press). Lithofacies assemblage I represents the type section of the Polly Brook Formation and assemblage II, the Leamington Member of that formation. Assemblage III (including the bivalve-bearing association III B) comprises the Joggins Formation in the vicinity of Springhill. The Springhill coalfield is designated as the type area for the Springhill Mines Formation, represented by the coal measures of assemblage IV. Assemblage V, a member of the Springhill Mines Formation transitional to currently unassigned beds of assemblage VI, is correlative with strata exposed near MacCarron's River, south of Joggins (Calder and Bromley, 1988). These strata have been assigned to the MacCarron's Creek Member of the Springhill Mines Formation by Ryan *et al.* (in press). The unassigned beds of assemblage VI are similar to essentially non coal-bearing beds in the Salt Springs and Mount Pleasant (Oxford) areas northeast of Springhill, which may be laterally transitional to, or underlie, the Ragged Reef Formation.

2.1 DISCUSSION OF THE CONCEPT OF FACIES

The application and interpretation of the term "facies" has varied widely since its first use in a geological sense by Steno in 1669 (summarized in Teichert, 1958). Gressly (1838) re-introduced the term and applied it to summarize the entire lithological and paleontological aspects of a stratigraphic unit, regardless of its stratigraphic position. Moore (1949), however, used facies to describe areally restricted parts of a stratigraphic unit, not unlike a member in formal stratigraphic nomenclature. It is this usage of the term that Shaw (1951) and Copeland (1959) employed in subdividing the Cumberland Group, as described in the preceding chapter.

The term facies in sedimentology should define a sedimentary product in an objective manner, avoiding the subjective interpretation of environment of deposition as a facies criterion (Middleton, 1978; Reading, 1978). Such interpretations can be made subsequent to the definition of a particular facies, after taking into account its relationship to associated facies. Miall (1984) defined a lithofacies as "a rock unit defined on the basis of its distinctive lithologic features, including composition, grain size, bedding characteristics and sedimentary structures", whereas a biofacies is defined on the basis of fossil components (either body parts or traces). Miall suggested that colour, being subject to diagenetic change, should not be used as a primary basis for lithofacies definition. Reading (1978), on the other hand, excluded neither colour nor fossil content from his definition of facies:

> "In the case of sedimentary rocks, it is defined on the basis of colour, bedding, composition, texture, fossils, and sedimentary structures. "Lithofacies" should thus refer to an objectively described rock unit."

The author prefers the definition of Reading (1978), which takes colour and clastic fossil components into account, as these will ultimately aid in paleoenvironmental reconstruction. The use of fossil components in lithofacies definition is considered to be valid so long as they are viewed as lithological entities within the rock, rather than as time-related biological communities. Miall (1984) interpreted a facies as being representative of an individual depositional event, whereas Reading (*ibid.*) once again made a broader interpretation, writing that a facies "should ideally be a distinctive rock that forms under certain conditions of sedimentation, reflecting a particular process or environment." Accordingly, emphasis in this study is placed on objective description without advance interpretation of depositional process.

2.2 LITHOFACIES NOMENCLATURE

Perhaps the most widely used lithofacies nomenclature scheme for continental and in particular fluvial rocks is that of Miall (1978). In light of the wide use of Miall's facies codes, in this study similar codes will be employed where applicable in describing the lithofacies of the Springhill coalfield. In certain cases these codes have been found to be too broad. This is most noticeable in the case of interbedded siliciclastic lithofacies and organic (coal) lithofacies. A further complication encountered in the present study is the necessary reliance on drill cores. Such cores generally do not reveal the meso- to macro-scale bedforms often considered to be diagnostic in the lithofacies designations of Miall and subsequent workers. The shortcomings cited above and the difficulties caused by reliance upon drillcore description have been addressed to a large degree by Le Blanc Smith (1980).

2.2.1 Lithofacies Groupings

Lithofacies of the Springhill coalfield (Table 2.1) have been divided into five groups, designated by the capital letters G,S,F,L, and C. Lithofacies group G (to denote gravel) of Miall (*ibid.*) includes conglomeratic strata. Group S (to denote sand) of Miall (1977) includes sandstones and coarse siltstones. Although Miall did not differentiate siltstones, the author has observed that coarse siltstones tend to exhibit sedimentary structures similar to some fine-grained sandstones whereas fine siltstone and claystones tend to show affinity. Group F (to denote fine-grained) of Miall (*ibid.*) includes the mudrocks: claystone, siltstone and mudstone. Grain-size criteria for the sedimentary rocks described in this study are those of the Udden-Wentworth scale. Coal has been elevated from the single facies of Miall (1977, 1978) to lithofacies group C whereas lithofacies group L (to denote limestone) incorporates carbonate facies, following the approach of Le Blanc Smith (1980).

2.2.2 The Integration of Coal Lithotypes in a Lithofacies Study

Few studies of the sedimentology of coal-bearing strata have treated coal as anything other than a single lithofacies that includes the entire spectrum of coal to carbonaceous shale (e.g. lithofacies C of Miall, 1977; Rust, 1978). Le Blanc Smith (1980) suggested finer sub-divisions of coal in lithofacies terms but did not implement them in his study. The definition of coal lithotypes on the basis of objective macroscopically discernible lithology (Stopes, 1919; ICCP, 1963) makes this scheme readily amenable to incorporation in a lithofacies scheme.

Coal has been subdivided in this study into four macroscopically recognizable lithofacies based on a) the distinction between humic and sapropelic coals and b) between these and "impure" coal, which comprises greater than 30% inorganic matter (ash) by weight and is usually interlaminated with mudrock (Table 2.1). The four lithofacies are: i) banded humic coal (Ch); ii) homogeneous, sapropelic coal (Cs); iii) impure coal comprising microbanded to interlaminated humic coal and

Lithofacies	Subfacies	Distinguishing Attributes	Interpretation of Depositional Process and/or Bedform
Gt		trough cross-stratified	tractive deposition of gravelly dunes in channels
Gg		crudely stratified; graded	sheetflow deposition
Gmf		massive, framework-supported	Gt, Gg described in core only
Gi		graded, thinly interbedded conglomerate and sandstone	episodic waning flow deposition of sand and gravel in traction and suspension
St		large scale trough cross-stratified	tractive deposition from dunes
Sp		planar cross-stratified	in-channel bars (in core may also be dunes-St)
Sr		undifferentiated ripple cross-stratified	ripple forms under lower flow regime
Sl		broad, low-angle cross-stratified	deposition in broad, shallow scours under low to high velocity
Sh		horizontally stratified	planar bed flow deposition of upper or lower flow regime
	Shl	horizontally laminated	upper or lower flow regime
Sg		poorly stratified and sorted coarse sand; often normal graded	waning distal sheetflow deposition from suspension, minor traction
Ss		graded, pebbly sandstone with with common plant stems	waning upper flow regime deposit overlying deeply scoured surfaces
Se		intraclasts	bedload deposits from eroded channel banks
Sm		massive	grainflow; liquefaction; rapid deposition of homogeneous sand
Sd		deformed stratification	original stratification obliterated by various causes
	Sdc	convolute lamination, load casts, injection features	deformation by liquefaction promoted by rapid deposition of heterolithic sediments
	Sdb	bioturbation	deformation by invertebrate burrowing (Sdbb) or rooting (Sdbr)
Si		interstratified sandstone with lesser mudrock	episodic overbank or drape deposition
Fl		laminated fines	episodic deposition of fines from suspension
	Flrh	rhythmic lamination	regular, episodic (?seasonal) deposition from suspension
Fm		massive	rapid deposition of mud under quiescent flow conditions

Table 2.1 Summary Table of Lithofacies of the Springhill Coalfield.

Lithofacies	Subfacies	Distinguishing Attributes	Interpretation of Depositional Process and/or Bedform
Fg		scattered coarse sand to fine pebbles in mudrock	dilute mudflow; overbank deposition from suspension of poorly constrained streams
Fi		interstratified mudrock with lesser sandstone	episodic overbank or drape deposition
Fcl		crudely micro-stratified carbonaceous matter	depositon of mud and (aquatic?) organic matter in ponds of fluctuationg level
Fc		carbonaceous matter common to abundant	mud and plant matter preserved under quiescent, reducing conditions
Csi		interlaminated sapropelic coal and mudrock (may include cannel shale)	subaquatic deposition of mud and finely detrital plant remains
Ci		interlaminated humic coal and mudrock (includes the terms "coaly shale" and "shaly coal", which are based upon relative proportions of organic to inorganic constituents	peat accumulation with contemporaneous siliciclastic deposition from muddy flood waters
C		humic coal (banded)	accumulation of peat from various plant types and parts under oxic to anoxic conditions
	Cf	fusain	partially oxidized plant matter
	Cd	durain	oxygenated subaquatic organic ooze deposits with possible corprogenic processes
	Cv	vitrain	primarily woody plant parts derived from humic-acid substances in peat; coalfied bark of drifted plant stems
	Cc	clarain	alternating to simultaneous accumulation of durain and vitrain under lower groundwater level than that for durain
Cs		sapropelic coal (nonbanded)	subaquatic deposition of finely detrital plant remains
Lc		aphanocrystalline with bounding coaly laminae	precipitation of $CaCO_3$ in organic-rich ponded water
Lcw		permineralized, fusainous wood	replacement of water-logged wood by $CaCO_3$
Lcc		cone-in cone structure	precipitation in shallow ponds with later diagenetic crystal growth
Lm		massive, muddy to finely arenaceous	precipitation in shallow flood plain ponds
Lb		fossiliferous packstone with bivalves	low energy deposition of fresh-water invertebrate remains and mud
Lp		granular to nodular texture, may be rooted	pedogenesis

Table 2.1 Summary Table of Lithofacies of the Springhill Coalfield (cont'd).

mudrock (Ci); and similarly interstratified sapropelic coal and mudrock (Csi). The microscopic lithotypes of humic coal (vitrain, clarain, clarodurain, durain and fusain) (International Committee for Coal Petrology, 1963) are herein assigned the status of subfacies of humic coal (lithofacies Ch); an alternate scheme would be to assign lithofacies status to the individual lithotypes. Sapropelic coal, which is rare in this study area, is not differentiated in this study due to the difficulty in macroscopic differentiation between cannel and boghead coals (Teichmüller, 1982).

2.3 LITHOFACIES ASSEMBLAGES, ASSOCIATIONS, SEQUENCES AND CYCLES

Lithofacies tend to exhibit an association with other lithofacies rather than a random occurrence throughout the basin-fill or stratigraphic record. Such naturally occurring groups of lithofacies may be referred to as lithofacies associations or assemblages. Potter (1959) defined a facies association as "a collection of commonly associated attributes". These attributes were given as "gross geometry (thickness and areal extent); continuity and shape of lithologic units; rock types .... sedimentary structures, and fauna (types and abundances)." Commenting on this definition, Miall (1984, p. 139) noted that "a facies association (or assemblage) is therefore based on observation; perhaps with some simplification." In this study, the term lithofacies association is used uniformally to describe relationship between individual lithofacies (e.g. sideritic mudrock -coal association). Lithofacies assemblage in this study is used to define groups of lithofacies at a scale similar to a member in lithostratigraphic nomenclature but without consideration of areal distribution, in keeping with recent usage in sedimentological studies of Nova Scotian coal basins (e.g. Rust *et al.*, 1984; Gibling and Rust, 1984; Gibling and Rust, 1987; Rust *et al.*, 1987).

The principle that a natural affinity exists between certain lithofacies within the stratigraphic record, was first stated by Johannes Walther (1894) in his "Rule of Succession of Facies" which states, in part:

> "The various deposits of the same facies area and, similarly, the sum of the rocks of different facies areas were formed beside each other in space, but in a crustal profile we see them lying on top of each other ..." (translated by Middleton, 1973).

Walther, in considering the succession of facies in the vertical profile was in effect describing what is commonly called a facies sequence. A facies sequence is defined by Reading (1978) as "a series of facies which pass gradually from one into the other". An erosive contact or hiatus commonly bounds the sequence. Heward (1978) applied specific criteria (especially thickness) to the term sequence (m -10's m thick) and also proposed the terms megasequence (10's - 100's m thick) and basin-fill sequences (100's - 1000's m thick). Megasequences comprise sequences and related beds, and basin-fill sequences embrace sequences and megasequences. His term megasequence is considered to be analogous either to large-scale facies sequences or lithofacies assemblages of this study. A sequence that occurs repeatedly in the vertical section is said to be cyclic, a concept introduced by Dawson (1854) in his study of coal-bearing strata exposed at Joggins. In its simplest form, a cyclic sequence involves repetition of a couplet, say AB, in the manner ABABAB etc. Duff *et al.* (1967) favoured synonymity with the term cycle.

Given Walther's law of facies, it follows that repetitive shifting of similar environments through time will result in cyclic recurrence of lithofacies sequences, although external (allogenic) sudden depositional and erosional events may interpose in the regular migration of facies belts (Anderson and Goodwin, 1990). Cyclic sequences are common in fluvially-dominated coal-bearing strata. The study of cycles within the so-called "coal measures' has formed the bulk of such investigations in cyclic sedimentation, particularly throughout the middle decades of this century. Wanless and Weller (1932, p. 1003) introduced the term cyclothem "to designate a series of beds deposited during a single sedimentary cycle of the type that prevailed during the Pennsylvanian period". Perhaps not surprisingly, this rather broad definition can be applied to cyclic sequences within many late Carboniferous coal-bearing fluvial sequences. The reader is referred to Duff *et al.* (1967) for discussion, in depth, of the subject of cyclic sedimentation; and to McCabe (1984) for an historical overview in specific relation to coal studies.

The lithofacies assemblages defined in this study of the Springhill coalfield typically exhibit a maximum thickness in the order of 200 - 500 m. Within the study area, six lithofacies assemblages have been identified: I) conglomerate assemblage; II) poorly-sorted thinly bedded assemblage; III) thinly bedded sandstone/siltstone assemblage, including the bivalve-bearing association IIIBv; IV) grey mudrock/multistorey sandstone assemblage;; V) variegated mudrock/sandstone assemblage; and VI) reddish mudrock/lithic sandstone assemblage. The major coal seams of the Springhill coalfield occur within assemblage IV. Accordingly, this study emphasises this lithofacies assemblage and its interaction with other assemblages with which it is in direct contact.

The lithofacies assemblages were seldom defined by the occurrence of a unique lithofacies but rather on the basis of lithofacies thickness, lateral and vertical relationships between lithofacies (particularly within individual cycles or "sub-assemblages"), the nature of bed contacts, texture (especially sorting), colour, and relative proportions of lithofacies and lithofacies groups (e.g. S:F, G:S). In rare instances, specific lithofacies were found to be restricted to only one assemblage; more frequently, common attributes of several lithofacies, such as the nature and occurrence of siderite and degree of sorting, were found to be useful tools in defining assemblages.

Lithofacies assemblages thus defined on the basis of a holistic sedimentological approach became apparent in the course of mapping and core logging (Calder, 1980; 1984b; 1986) but were more precisely characterized by subsequent quantitative assessment of measured sections (Appendix A). Due to the generally sporadic exposure of strata within the study area, drillcore was utilized where available in this study to obtain data on lithofacies abundances. The extrabasinal conglomeratic assemblage, however, was characterized through compilation of mapped outcrop sections within and beyond the study area due to minimal drillcore coverage. Lithofacies abundance data for the various assemblages are given in Appendix A, and summarized graphically in Figure 2.1.

This chapter deals solely with the sedimentological description and interpretation of the various assemblages. Spatial and other stratigraphic considerations are discussed in the following chapter . For the assistance of the reader, the geographic occurrence of the assemblages within the study area is given in this chapter.

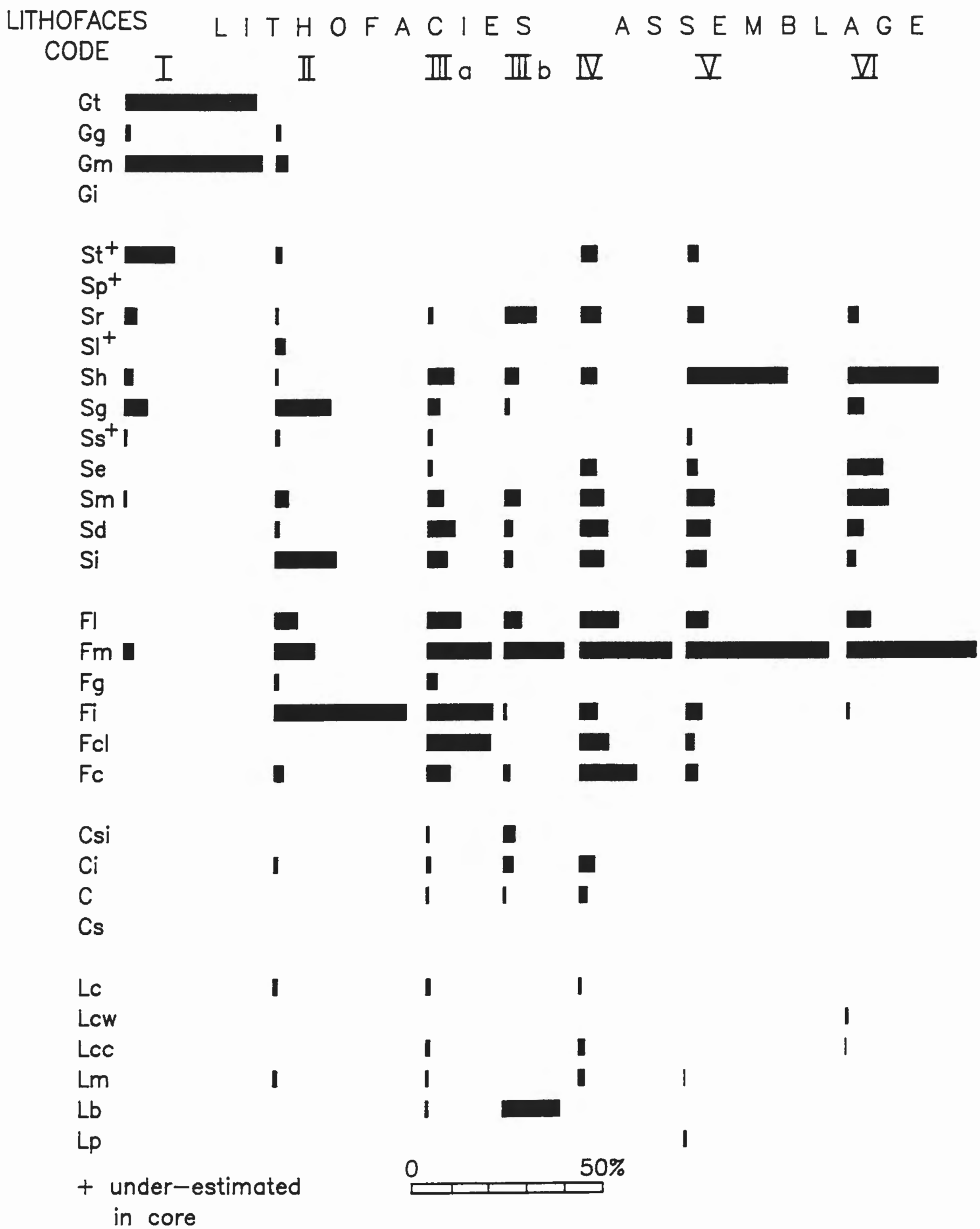


Figure 2.1 Relative abundance of lithofacies in assemblages of the Springhill coalfield.

2.4 CONGLOMERATE LITHOFACIES ASSEMBLAGE (I)

2.4.1 Occurrence:

Data pertaining to the conglomerate lithofacies assemblage are derived entirely from outcrop descriptions, as the assemblage is not represented in existing drillcore. Within the study area, the conglomerate assemblage is best exposed on Polly Brook in the southeast. The measured section is depicted in Figure 2.1. In the northeast, within the Salt Springs coal district, the conglomerate assemblage is also exposed along Black River on the northern flank of the Black River diapir.

Outside the study area, the conglomerate lithofacies assemblage is well exposed along a 7 km length of Polly Brook east toward the village of River Phillip. Southwest of the map area, within the Athol Syncline, the assemblage is exposed on numerous northerly-flowing streams draining the northern flank of the Cobequid Highlands, such as one running along the east side of Lynn Mountain road, and Fordyce Brook east of Newville (or Halfway) Lake.

Within the Athol Syncline, the conglomerate assemblage is also partially represented in drillcore from Cera-Caledonia hole CC74-3 and NSDME drillhole SB-3. Drillhole CC74-3 at Southampton 11 km west of the map area and 20 km west of the Polly Brook section, intersected strata of the conglomerate lithofacies assemblage in the lowermost 240 m of the drillhole. SB-3 was drilled at Styles Brook, 5 km north of the map area and 9 km north of the Black River section, at the eastern extremity of the Joggins-Chignecto coalfield, on the northern limb of the Athol Syncline.

2.4.2 Description:

The sporadic exposure along the streams undoubtedly renders an incomplete picture of the conglomerate lithofacies assemblage and may skew the data in favour of lithofacies more resistant to weathering. The conglomerates are polymictic, with pebble to cobble-size clasts up to 18cm in diameter (Plate 2.1). Trough cross-stratified conglomerate (Gt) and massive to crudely stratified conglomerate (Gm), both with sandy to granule-sized matrix, are the most abundant lithofacies (Figure 2.1). Together they comprise 70.3% of the exposed part of the section. Sandstone lithofacies (S) account for 26.4% of the section. Mudrocks (grey siltstone) are rare, comprising a mere 2.2% of the total. Coal and limestone lithofacies are not present in the measured section. The ratio of conglomerate lithofacies (G) to finer-grained lithofacies (S + F) varies from 2.5:1 on Polly Brook to 3:1 on Black River.

Plate 2.1 Polymictic pebble to cobble conglomerate of lithofacies assemblage I exposed on Halliday Brook, south-west of the Springhill coalfield.

2.4.2.1 Large scale sequences

The estimated 250 m-thick stratigraphic succession exposed on Polly Brook (Figure 2.2) can be divided into three sequences (cf. megasequences of Heward, 1978) which appear to exhibit systematic grain-size trends. The lowest sequence, from 0 to approximately 124 m, generally fines upward. Pebble to cobble conglomerate prevails in the lowermost 44 m of the section, giving way to predominantly fine pebble conglomerate to 112 m, and thence to sand-dominated strata to 124 m. The maximum clast dimensions within this sequence confirm the trend. A coarsening, then fining-upward sequence occurs from 124 m to 150 m +. Medium-grained sandstone (St) grades through granule and locally pebble conglomerate in the basal 1 - 7 m and gives way to a relatively consistent pebble-rich conglomeratic section from 131 - 135 m. Above 146 m another sand-dominated succession appears. The upper sequence, from 225 m to 250 m, shows no clear grain-size trend. The generally consistent pebble conglomerates do, however, show some evidence of upward-fining to granule to pebble-size fractions above 240.7 m.

2.4.2.2 Smaller scale sequences:

Smaller scale lithofacies sequences can be readily identified within the Polly Brook section. At the base of the measured section, trough cross-stratified conglomerate is overlain by thin beds of normally-graded, fine to coarse-grained sandstone exhibiting trough cross-stratification (St). At 41 - 43.5+ on the measured section, steep-sided erosion surfaces (scours) greater than 0.60 m deep occur within matrix-supported, trough cross-stratified conglomerate (Gt). Rare, thin lenses of conglomeratic sandstone overlie shallower scoured surfaces at the top of the units of Gt. From 240 - 241 m (Figure 2.2), matrix - to clast-supported cobble conglomerate (Gt) with crude imbrication occurs in a bed greater than 0.75 m thick, and is overlain by coarse granule to pebble matrix-supported conglomerate (Gm) in 2 cm-thick beds that are possibly low-angle cross-stratified. The bases of most conglomerate-rich cycles are erosional where visible, as at 124.2 m on the measured section (Gt overlying St).

A coarsening-, then fining-upward sequence dominated by sandstone lithofacies is evident in the section from approximately 122 - 123.9 m. This sequence comprises the following series of lithofacies: 1) an erosive base overlain by medium-grained, trough cross-stratified sandstone (St) with *in situ* plant stems; erosively overlain by 2) normal-graded very coarse-grained sandstone (Sg); overlain (uncertain contact) by 3) a similar unit of St to that described above; in turn erosively overlain by 4) cross-stratified sandstone (Sr to St: set size approximately 5 cm) with an upward increase in

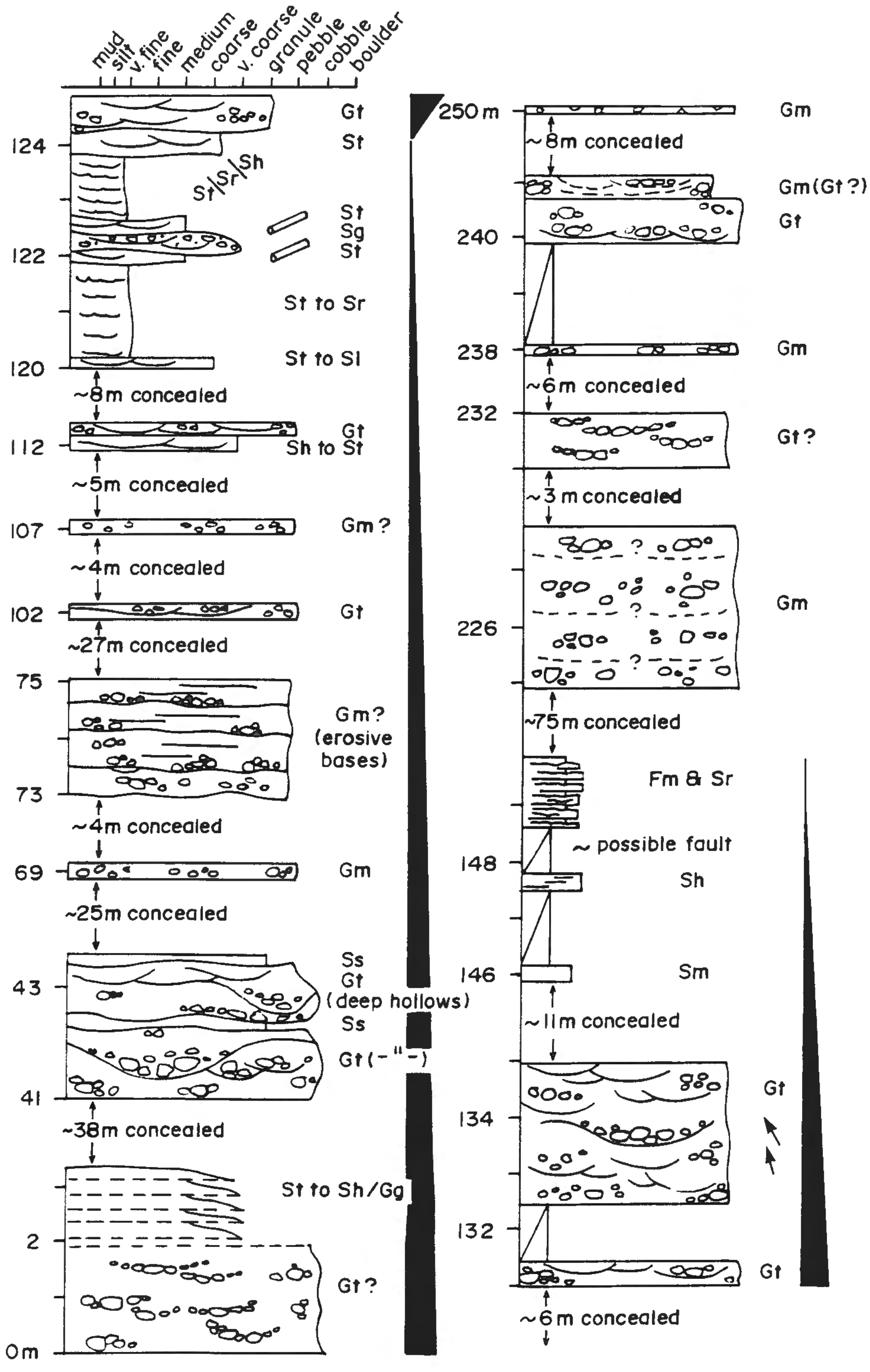


Figure 2.2 Measured section, conglomerate lithofacies assemblage (I), south Polly Brook.

wavy-bedded sandstone (Sr to Sh) and concomitant decrease in grain size. In summary, sequences are relatively thin, in the order of a few metres thick, and exhibit systematic upward changes in lithofacies, grain size, and sedimentary structures.

2.4.3 Interpretation:

The conglomerate lithofacies assemblage exposed within the map area is interpreted as a fluvial succession, in part on the basis of the general abundance of cross-stratification and erosional contacts. Preserved lithofacies within exposed cycles indicate that bedload transport predominated during active periods of flow, after which flow waned rapidly. Trough-cross-stratified conglomerate within these cycles probably formed by migration of gravelly dunes during high-stage flow within channels (Rust, 1978), in a manner similar to bedforms described by Fahnestock and Bradley (1973) and Galay and Neill (1967). Massive to crudely stratified conglomerate sequences are interpreted as low-relief bedforms, probably longitudinal bars (Hein & Walker, 1977; Rust and Koster, 1984). Upward-fining longitudinal bars are well described from braidplains or rivers (as in references cited) but occur as well on alluvial fans (Kesel and Lowe, 1987; Darby *et al.*, 1990 and observations by the writer in Death Valley). Such beds are generally less regular and may imply more episodic accumulation.

Bar deposits were subsequently modified in at least two manners: 1) lower portions were locally scoured to depths of at least 0.60m by episodic surges in stream power as at 41-43m and 134m (Figure 2.2); and 2) upper portions were subject to modification by waning flows which scoured shallow hollows infilled by cross- to horizontally stratified sands, as at 0-3m and 41-43.3m (Figure 2.2).

The abundance of gravel and the nature of the sequences described above is indicative of braided river deposition (Rust, 1978; Miall, 1978; Rust and Koster, 1985), but also of alluvial fan deposition (Kochel and Johnson, 1984; Kesel and Lowe, 1987; Darby *et al.*, 1990). Trough cross-stratified gravel is a significant lithofacies within distal, braided rivers (Rust, 1978) such as the mid-reaches of the Donjek River, Yukon (Miall, 1977). However, massive to crudely stratified gravels are common in proximal braided rivers. Some of the apparent horizontal stratification observed in outcrop could in fact be low angle cross-stratification.

It is uncertain, therefore, on a strictly sedimentological basis whether the assemblage formed part of an alluvial fan proper or of an extensive braidplain. The predominance of braided fluvial deposits in the examined section cannot be considered conclusive evidence of a braidplain. Both humid-zone and arid-zone alluvial fans can be relatively rich in water-laid gravel deposits (Bluck, 1964; Hooke, 1967; Bull, 1977; Schumm, 1977; Kochel and Johnson, 1987; Kesel and Lowe, 1987). This largely geomorphic question, which involves provenance, paleocurrent and regional distribution patterns, is therefore discussed subsequently in Chapter III.

Large-scale (mega)sequences are poorly defined, but presumably reflect long-term variation in subsidence rate or climate (cf. Steel, 1976; Steel *et al.*, 1977; Blair, 1987; Blair and Biloudeau, 1988; Frostick, 1989). Small-scale sequences reflect channel fills, either due to bar formation and migration or to flood aggradation (cf. Hein and Walker, 1977; Gustavson, 1974; Miall, 1977). The thinness of such sequences (a few metres) is typical of alluvial fan and braided facies but especially of fans (e.g. Bull, 1972).

2.5 POORLY SORTED, GRADATIONALLY INTERBEDDED LITHOFACIES ASSEMBLAGE (II)

2.5.1 Occurrence:

The poorly sorted, thinly interbedded lithofacies assemblage crops out in the south half of the map/study area, in and about the Springhill Syncline. Rapid lateral changes in lithological character and interfingering with other lithofacies assemblages renders accurate thickness determination very difficult. On Black River 1.2 km northeast of Black River Road, Rodney, the poorly sorted lithofacies assemblage is intercalated with both the conglomerate and mudrock/sandstone lithofacies assemblages over a stratigraphic interval of 290 m. In drillhole CC74-3, 20 km to the west, the poorly sorted assemblage is only 44 m thick, which indicates that it is better developed in the east half of the Athol Syncline.

2.5.2 Description:

The poorly sorted lithofacies assemblage has been partially cored in numerous drillholes along the axis and south limb of the Springhill Anticline. The relative abundance of lithofacies in one such section (from drillhole SH38; 120.11 - 122.0 m, and 124.0 - 162.20 m) is given in Appendix A. The most abundant lithofacies is Fi, which alone comprises 40.9% of the section; the two interstratified sandstone/mudrock lithofacies (Fi and Si) account for greater than 50% of the strata. Conglomerate (G) comprises less than 4% of the section, but granule to pebbly sandstone is common (15.9%). Coals are present but rare and impure. Limestones are only slightly more abundant in the section, comprising 1.2% of the total.

The abundance of thinly interbedded sandstone and mudrock (chiefly siltstone) is one of the chief characteristics of the assemblage. Further characteristics include the generally ill-sorted texture and poorly developed stratification of the coarser sedimentary rocks (Plate 2.2). Beds of this assemblage are commonly relatively well indurated.

The sandstone : mudrock ratio is of little use in characterizing the poorly sorted assemblage due to the great abundance of thinly interstratified mudrock and sandstone which may be assigned to either lithofacies Fi or Si depending on local proportions of the two constituents. The lithologic dissimilarity between the conglomerate assemblage and the poorly sorted assemblage is obvious, however, when considering the G: S + F ratio. The ratio is 2.5:1 for the conglomerate assemblage on Polly Brook but only to 0.04:1 for the poorly-sorted assemblage in drillhole SH38.

Some insight into larger scale stratification and lithofacies relationships is provided by exposures, albeit scarce, within the map area. One such section, which outcrops along the Wyndam Hill road 2 km east of Leamington, is depicted in Figure 2.3. A summary of lithofacies abundance for this 7 m section (Appendix A), shows that 11.4% of the measured section is composed of poorly stratified conglomerate. Interstratified lithofacies Si comprises 49.2%, and cross-stratified sandstones account for 22.4% of the total. Mudrock (5.7%), coal and limestone lithofacies (nil) are probably under-represented in this exposure.

Stratification within the various lithofacies is chiefly horizontal to gently undulating; trough cross-stratification is poorly developed. Sandstone lithofacies commonly exhibit normal grading, although one bed 0.46 m thick comprises lithofacies Sg (coarse grained) coarsening upward to Gg (fine granule). Coarse-grained sandstone locally exhibits low-angle cross-sets (Sl) decimeters thick by metres wide (width:thickness ratios of approximately 10:1); (5.6-5.7 m position in Figure 2.3). The cross-sets commonly exhibit internal cm-scale, near-horizontal stratification and have low-angle, convex-upward erosional basal surfaces.

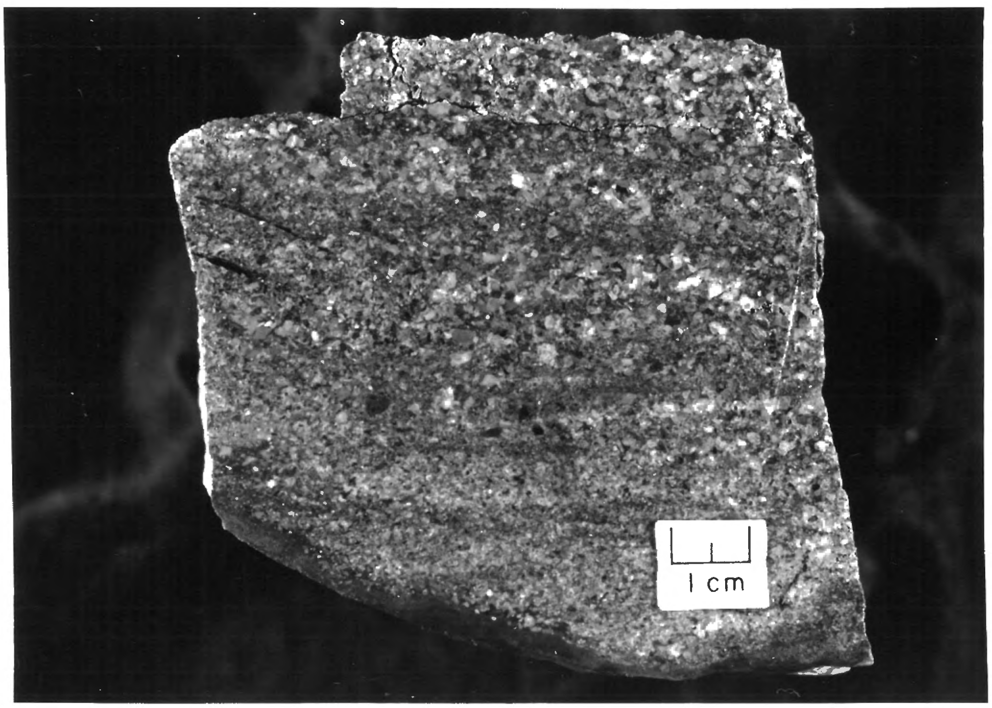


Plate 2.2 Feldspathic granule conglomerate exhibiting crude horizontal stratification, poorly sorted lithofacies assemblage II.

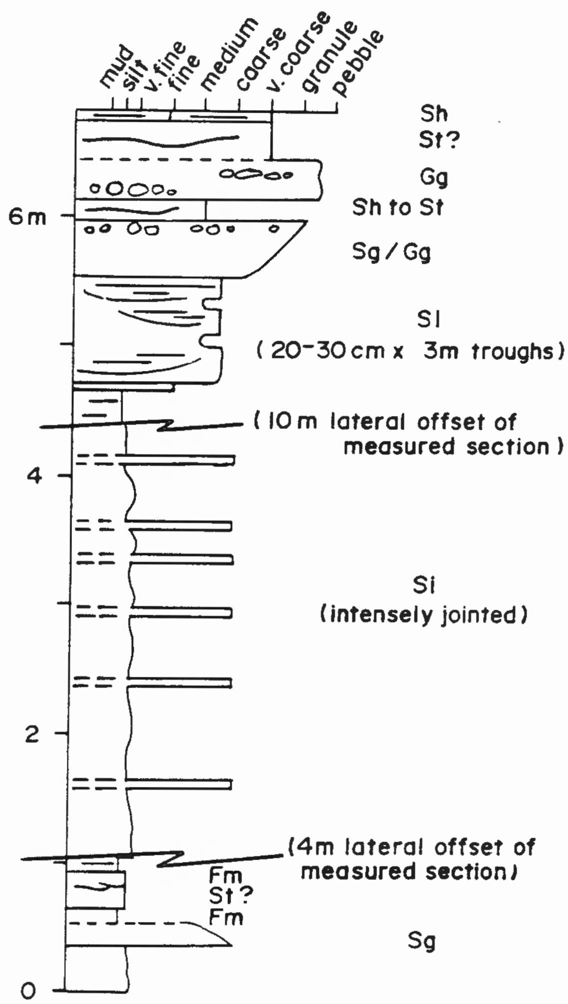


Figure 2.3 Summary vertical section, poorly sorted lithofacies assemblage (II) exposed on Wyndham Hill road between Leamington and Rodney.

The presence of normal graded, crudely stratified sandstone (Sg) forms one of the most characteristic attributes of the assemblage. The facies is commonly feldspathic and poorly sorted. Lithofacies Sg grades locally to the finer grained Fg or coarser grained Gg. Although lithofacies Sg forms only 15.9% of the assemblage as compared to 40.9% for lithofacies Fi, much of the interstratified sandstone component of Fi comprises poorly-defined, graded Sg.

In this assemblage, lithofacies Sg occurs in gradationally interstratified deposits which form irregular laminae and lenses and beds from 3 - 30 cm thick; erosional lower contacts are rare. Feldspathic sand- to granule-sized grains in places appear as scattered grains within silty mudrock (lithofacies Fg). Some lenses are up to 40 cm wide with a thickness of only 3 cm (13:1 ratio). Sg also occurs as isolated strata in repetitive, normal-graded cycles of 1 - 3 dm thickness or as amalgamated cycles which result in the general appearance of a crudely laminated coarse sandstone or granule conglomerate. A few cycles grade upward to fine-grained, ripple cross-stratified sandstone (Sr) (Plate 2.3). Rootlets are commonly present within the mudrocks.

2.5.3 Interpretation:

The assemblage is considered in general terms to be of fluvial origin; modification by pedogenesis is also evident. The assemblage is lithologically transitional to both the conglomerate and mudrock/multistorey sandstone assemblages.

Several lithofacies are interpreted as having formed within fluvial channels. These include: trough and low angle cross-stratified sandstone (St and Sl, respectively); sandstone with mudrock intraclasts (Se); and certain conglomerate lithofacies which overlie erosional surfaces. Thick, well-defined channel deposits, however, are rare. This suggests that in-channel deposition prevailed only at discrete intervals and locations and that channels occupied specific tracts for relatively short periods of time. Well-established channel tracts were, therefore, only rarely formed, flow having been ephemeral. The marked variation in grain size of in-channel deposits is evidence of variable stream power.

In contrast, sediments interpreted as sheetflow deposits are abundant. These include graded, crudely stratified and ill sorted, pebbly mudrock (Fg), sandstone (Sg) and conglomerate (Gg), which are commonly interstratified with siltstone (Fi). These cm-dm rhythmic cycles indicate bedload transport followed by waning flow deposition of mud or less commonly ripple-stratified sand (Sr; Plate 2.3), and were formed by episodically fluctuating discharge; each cycle represents a single episode of

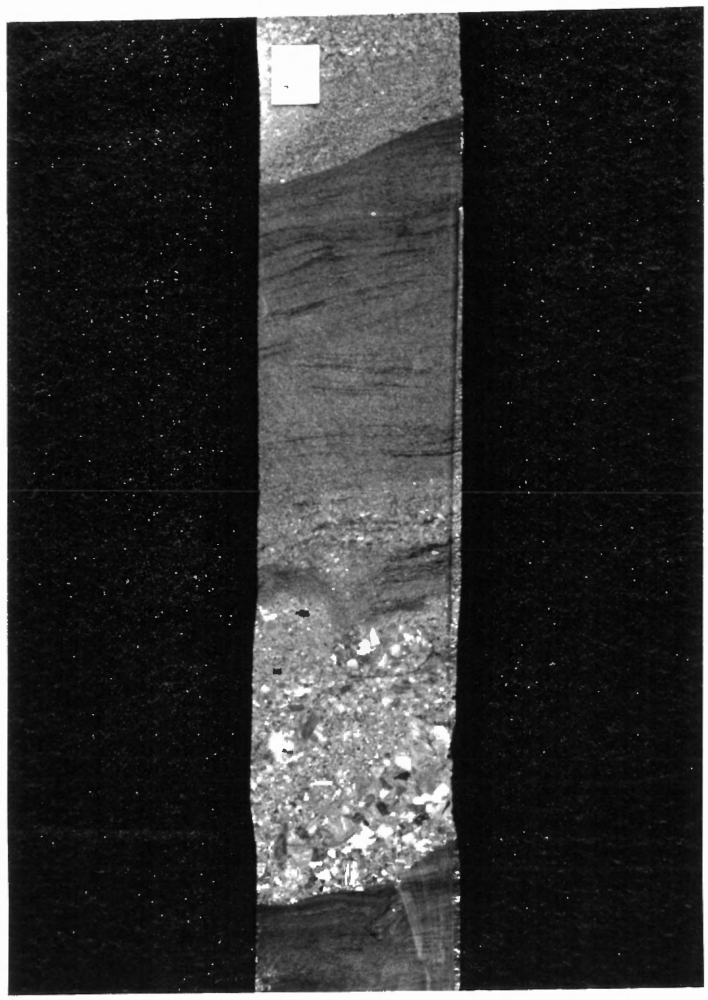


Plate 2.3 Erosionally based ill-sorted granule conglomerate fining upward abruptly to ripple cross-stratified sandstone, corehole SH38.

waning flow. Their origin is attributed to flashy, ephemeral sheetflow followed by rapid abatement; rainstorms may have been the triggering process. Although these sheetflow deposits could have formed within topographic channels, the deposits of modern, ephemeral sand-bed streams show few similar features (Karcz, 1972; Williams and Rust, 1969).

Vegetation, which either survived such flood events or became re-established between episodes, probably contributed to the poor development of stratification within these sequences. Vegetation may have retarded flow and thus aided sedimentation. On certain modern arid-zone alluvial fans this results in a feedback mechanism whereby vegetation benefits from the presence of the groundwater (Bull, 1977).

The crude lenticular stratification of graded pebbly sandstone may have resulted from deposition above local scours. Alternatively, strata may have formed as thin sand ribbons, such as the features described by Karcz (1967) within ephemeral flash-flood deposits of the Negev Desert. Their interpretation as bedforms is hindered by a general paucity of outcrop and virtual absence of three-dimensional examples.

The formation of poorly sorted mudrock containing feldspathic granules (Fg) suggests a rapid loss in stream competence. This may have resulted from a sudden spreading of flow from crudely developed, shallow (evidenced by Sl described earlier) channels onto the depositional surface, as on extreme distal areas of alluvial fans in Death Valley (Plate 2.4). The local change from channelized to sheetflow that occurs ephemerally on the Dead Mesquite fan of Arizona (Packard, 1974) provides a similar analogue. This channel to sheetflow interpretation for these lithofacies is favoured because it explains the transition from Gt/Gg to Sg and Fg. The thinly bedded lithofacies Sg and in particular Fg could also represent dilute distal mudflows (Table 2.1). Both modes of deposition could have been coeval, however, as on modern alluvial fans of Papua (Ruxton, 1970).

The sediments which formed this assemblage may have been deposited as a terminal fan lobe at or beyond an alluvial fan toe through processes similar to those described by Parkash *et al.*, (1983) for the modern Markanda terminal fan of India. The Markanda River flows ephemerally, depositing thin, single-storied sand bodies and overbank sediments following monsoonal triggering. Highly variable discharge results, and sediments are deposited rapidly, as waters infiltrate into the detritus of the "fan". Landforms such as these are now referred to as 'plains fans' (North *et al.*, 1989).

Plate 2.4 Shallow, discontinuous channels on distal alluvial fan, Tucki Mt., Death Valley.

In summary, the poorly sorted lithofacies assemblage II is interpreted as deposits of poorly constrained channel flow, sheetflow and possibly dilute mudflow. Based on the vertical distribution and nature of lithofacies sequences within the assemblage, highly variable deposition occurred, with ephemeral discharge probably triggered by heavy rainfall (cf. Karcz, 1967; Parkash *et al.* 1983). Abundant roots within interbeds of mudrock (reddish, mottled or grey) and within graded, poorly sorted Sg and Fg indicates that the fan-toe/bajada surface was at least in part vegetated. Pedogenesis was predominantly restricted to root bioturbation, and only rarely reached the type 1 caliche development of Allen (1974), calcrete having been observed rarely (e.g. drillhole SH20A).

2.6 THINLY INTERBEDDED LITHOFACIES ASSEMBLAGE (III)

2.6.1 Occurrence:

Within the Springhill coalfield, the thinly interbedded lithofacies assemblage is known only from drillhole SH74 where it attains a thickness of at least 175 m (Figure 2.4).

2.6.2 Description:

The assemblage is characterized by thinly interstratified beds of siltstone and fine sandstone with gradational contacts; erosive contacts are rare to absent. Mudrocks are commonly silty and indurated; claystone is rare. Mottled and reddish-brown hues are exhibited by 38 percent of mudrocks in the lowermost 49 m of the assemblage cored in SH74. The majority of the mudrocks of the measured section in drillhole SH74 exhibit a grey to slightly greenish-grey hue. Sequences of sandstone commonly coarsen, or coarsen then fine upward from gradational bases. Sandstones units are not erosively amalgamated, hence the assemblage does not appear to contain multistorey sandstone bodies. Sandstones invariably exhibit deformation structures and are commonly rooted. In places (for example, between 224.31 and 233.44 m, Figure 2.3) the assemblage appears transitional to the coal-bearing assemblage IV with the development of blacker, more organic- and clay-rich mudrocks and thin coals. Limestone occurs as isolated, spherical bodies several centimetres in diameter within mudrock. It is uncertain whether these bodies originated as concretions, perhaps around roots, or as detached load-balls ("pseudo-nodules").

The relative abundance of lithofacies from the only available reference section, i.e. SH74, is summarized in Appendix A. The assemblage is rich in mudrock (siltstone, primarily) which comprises

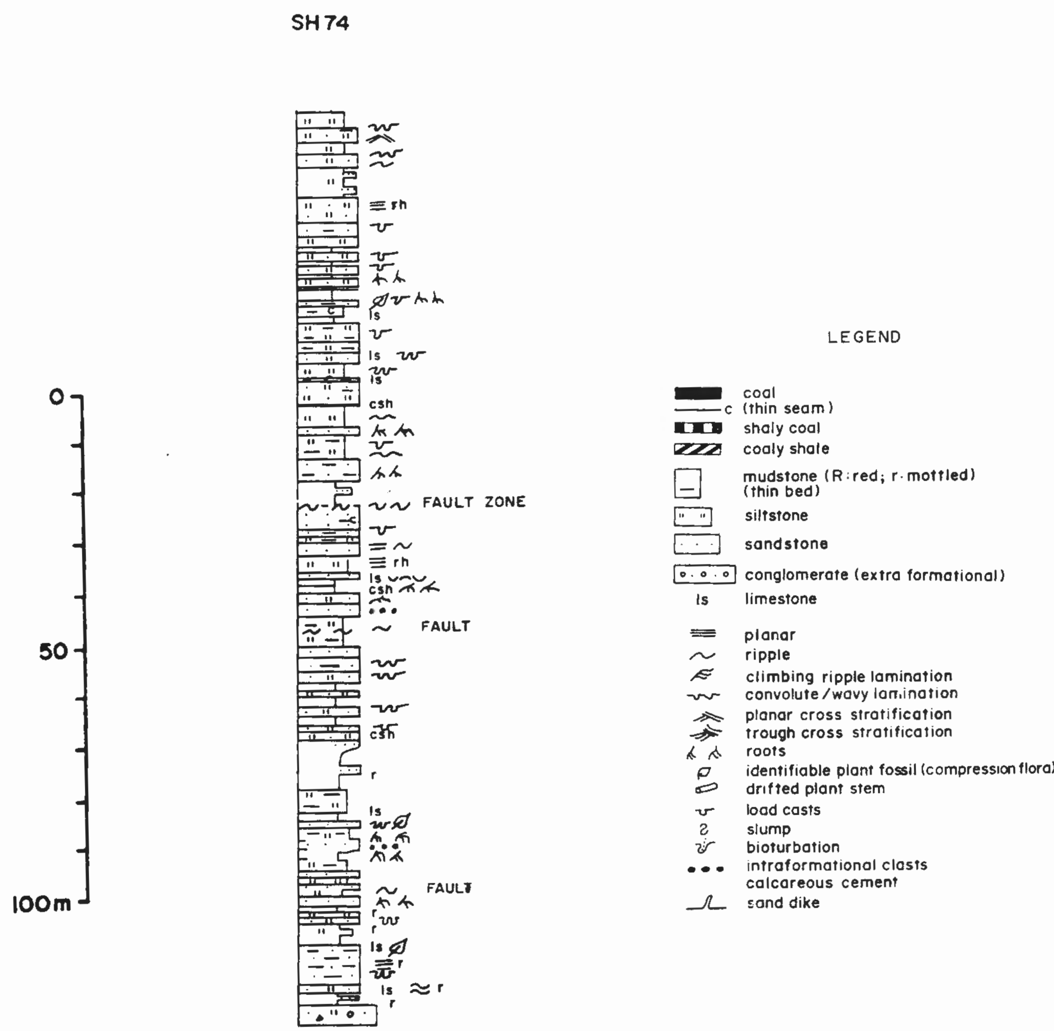


Figure 2.4 Lithofacies assemblage III, corehole SH74 (from Figure 3.1, in rear pocket).

approximately two-thirds of the strata. Sandstone comprises the bulk of the remainder, with limestone and coal lithofacies accounting for 1.9 and 1.2% respectively.

Although the assemblage is lithologically similar in many respects to the poorly sorted lithofacies assemblage II, certain key departures are evident (Figure 2.1). Several lithofacies are represented in the thinly bedded assemblage but are absent or uncommon within the poorly sorted lithofacies assemblage. These include sandstone exhibiting soft-sediment deformation (Sd); horizontally laminated sandstone (Shl); rhythmically laminated mudrock and sandstone (Flrh); mudrock that bears crudely, extremely thinly stratified carbonaceous matter (Fcl); carbonaceous mudrock (Fc); and thin coals and mudrocks of a quasi-sapropelic nature (Csi). Conglomerates are conspicuously absent from the assemblage.

2.6.3 Interpretation:

The deposits of this assemblage are interpreted broadly as a lake-fill (cf. "bay-fill" of Elliot, 1974 wherein the term bay has no implication with respect to openness to the sea) in part due to the thickly interstratified nature of the rocks and paucity of coals and multistorey sandstones. The rhythmic interbedding of fine sand and silt is attributed to "overbank" flooding processes (Coleman, 1969; Elliot, 1974). The lack of cross-stratification and of erosional boundaries within sandstones is evidence of unconfined flow. Such deposits can be deposited with or without breaching of channel banks: a) sediment-laden waters may top confining banks and form sheet flows (Coleman, 1969; Elliot, 1974), or b) distal crevasse splays may be deposited from density currents (Coleman, 1969). While both mechanisms involve unconfined flow, if the latter were the process that prevailed during deposition of the assemblage, crevasse channel-fills should be present as well and apparently are not. The nature of the contributing fluvial system will be discussed later.

Coarsening-upward and coarsening-upward then fining-upward silt and silty sand sequences in the lower Mississippi River valley occur in both distal levees and distal splay lobe deposits (Farrell, 1987). Farrell cited three possible interpretations for the formation of such sequences: 1) sheet flood deposition as an initial phase in channel belt development; 2) progradation of overbank sediments related to meander migration; and 3) progradation of a lacustrine delta.

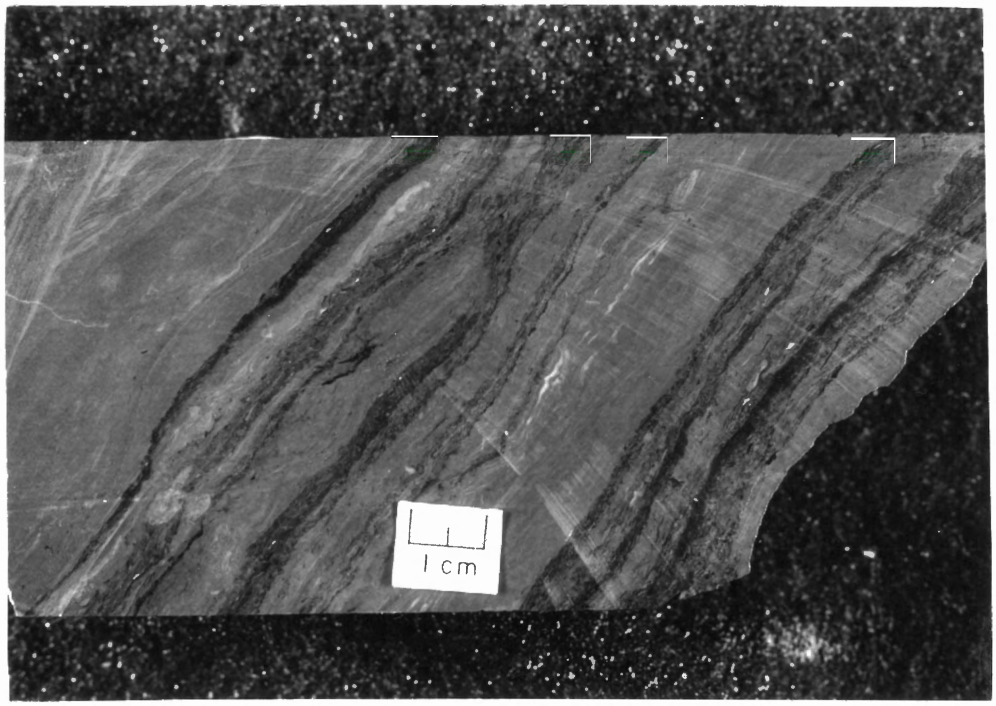
The strata show evidence of rapid deposition from suspension: load-cast, gradationally based sandstones, some capped by finely macerated carbonaceous debris and load-deformed thin interstrata of silt and fine sandstone are abundant. The interstratified nature of the association indicates episodic deposition. Minor development of ripples resulted from a basal traction carpet of the suspended load flow.

Mudrock and interstratified mudrock/sandstone containing crudely micro-stratified carbonaceous matter (lithofacies Fcl; Plate 2.5) is most abundant in this assemblage. Microscopic examination of polished core specimens under incident and blue-light excitation reveals that the "micro-laminae" are composed largely of detrovitrinite and semifusinite, with abundant liptinite including alginite, bituminite and detrital bitumen (Plate 2.6). The lithofacies indicates periodic accumulation of derived, ?biodegraded particulate plant matter. Possible modes of origin for this lithofacies include: i) hypautochthonous to allochthonous accumulation with derivation from waters or winds coursing through vegetated (peat?) lands; ii) sapropelic accumulation below a floating rhizome-network (cf. detrital peats accumulating below papyrus mats in modern Ugandan swamps: Lind, 1955; Beadle, 1974); or iii) a histosol soil horizon, the vestige of an oxidized, highly decomposed peat deposit. Lithofacies Fcl exhibits petrographic similarities to cannel shales of the Joggins Formation, as described by Gibling and Kalkreuth (1991). Vegetation was most certainly abundant during the deposition of this assemblage, and may have existed marginally to the site of accumulation. The greenish-grey and rarely mottled hue of the mudrocks of this assemblage is indicative of partially or periodically oxygenated waters of shallow and/or fluctuating levels (Coleman, 1966). Common roots within sandstone units is further evidence of either fluctuating or persistently shallow lake level.

2.7 BIVALVE-BEARING LITHOFACIES ASSOCIATION (III Bv)

2.7.1 Occurrence

Within the Springhill coalfield the bivalve-bearing lithofacies assemblage is known only from one drill core intersection, namely SH94, in the north of the coalfield. However, it makes a greater contribution to the basin-fill within the Joggins-Chignecto coalfield on the north limb of the Athol syncline, in the Salt Springs coal district to the northeast of Springhill and at Roslin approximately 20 km northeast of Salt Springs (Figure 1.2). At Salt Springs, the assemblage is exposed on the east bank of Black River 200m north of the confluence with Deep Brook. The occurrence of pelecypod-bearing strata is an important diagnostic criterion for the lithostratigraphic definition of the Joggins Formation (Ryan *et al.*, in press).



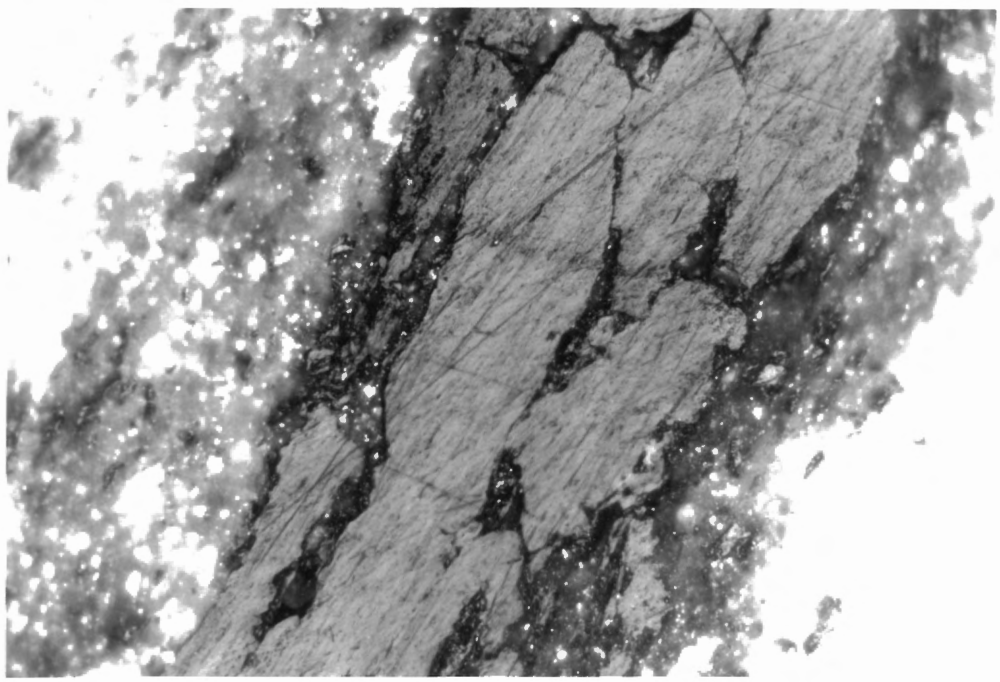


Plate 2.5 (Top:) Crudely laminated impure carbonaceous matter within greenish-grey mudrock (Fcl), lithofacies assemblage III, corehole SH74.

Plate 2.6 (Botton) Photomicrograph under incident (reflected) light (x 500) of impure carbonaceous lamina from lithofacies Fcl, corehole SH74. Lamina comprises degraded vitrinite/and lesser liptinite.

2.7.2 Description

As its name implies, this assemblage is defined primarily by the presence of limestone (packstone) or mudrock composed in part of shells of pelecypods and tests of ostracods (lithofacies Lb, Plate 2.7). In other respects, it is similar to lithofacies assemblage III and is described herein as an association of that assemblage. Other lithologically diagnostic attributes evident in drillcore from SH94 include even, rhythmic interlamination of fine sandstone and mudrock (Shrh or Flrh, depending on sandstone : mudrock abundance), and weakly banded (quasi-sapropelic) coal.

The bivalve-bearing strata described below from corehole SH94 may not be representative of the association elsewhere in the basin. Whereas this study focuses on the paleogeography and ecology of the ancient peat-forming systems of the Springhill coalfield, a comprehensive sedimentological analysis of these areas is beyond the defined scope of the research. It is nonetheless important to establish a broad, if only cursory, overview of the bivalve-bearing association, in light of its potential significance in basin analysis (Calder, 1981a, 1984; Calder and Naylor, 1985). Consequently, a brief qualitative summary of the bivalve-bearing strata in the basin follows, based on the author's observations unless otherwise indicated.

Throughout the Cumberland Basin, bivalve-bearing mudrocks and/or limestones are commonly underlain by humic coal seams. The bivalve-associated seams are invariably thin (<1 m) and high in sulphur (4.6-7.6%) in comparison with the major seams at Springhill (0.6-2.1% S; Copeland, 1959). The seams of the Joggins-Chignecto coalfield exhibit irregular thickness variation. Shales transitional from impure sapropelic coals are commonly found in close proximity to the bivalve-bearing lithofacies and exhibit characteristics of both impure sapropelic coal (cf. oil shale) and bioclastic bivalve-bearing limestone ('calcareo-bituminous shale' of Logan, 1845, and Copeland, 1959). Petrographic examination of a bivalve-bearing shale overlying the Forty Brine seam (coals 19-20, Division 4 of Logan, 1845) at Joggins (Gibling and Kalkreuth, 1991) indicated the bed to be composed largely of detrital humic matter, thereby falling within the classification of a cannel shale.

Fauna identified from bivalve-bearing lithofacies at Joggins, by J.E. Pollard and M.A. Calver (Duff and Walton, 1973) include species of the ostracod *Carbonita*; pelecypods *Curvirimula sp.*, cf. *Naiadites (Anthracomya) elongata; Naiadites longus*, the annelid worm *Spirorbis sp.* and fish remains represented by *Rhabdoderma sp.* Duff and Walton identified two types of bivalve-bearing limestones at Joggins: i) a pelecypod-bearing type, and ii) an ostracod-bearing type. *Anthraconauta sp.* have been tentatively identified from Salt Springs (Plate 2.7). Spiral tubes of the polychaete worm *Spirorbis* are often found attached to coalified compressions of fallen *Calamites* stems, which are the most commonly encountered compression flora within the assemblage.

An approximately 20 m thick overall coarsening-upward sequence of impure coal, fossiliferous limestone and mudrock; and rhythmically laminated mudrock and sandstone (Flrh to Slrh) was cored in drillhole SH94 (Figure 2.5). The upper leaf of the 1.74 m thick coal seam leaf is gradationally overlain by a generally coarsening-upward, 0.93 m thick sequence: comprising mudrock (Fm and Fi) 0.30 m thick overlain gradationally by a 0.63 m thick section of thinly bedded sandstone with soft-sediment deformation (Sd) thickly interlaminated with impure coal (Ci). This mudrock and interbedded sandstone/coal sequence is sharply overlain by a 1.40 m-thick section (Figure 2.5) of pelecypod and ostracod-bearing carbonate packstone (lithofacies Lb). Samples from this unit were submitted to B. Mamet, University of Montreal, for faunal analysis. Microscopic examination (B. Mamet, written communication, 1983) revealed fossilized remains of invertebrate (serpulopsid worm tubes and gastropods) and vertebrate (freshwater shark teeth and vertebrae) fauna. An overlying, generally coarsening upward section, 14 m thick, is composed primarily of rhythmically laminated mudrock and sandstone (Flrh to Slrh). This sequence is sharply (probably erosively) overlain by coarse-grained feldspathic sandstone (Sm to Sg) at the top of the section depicted in Figure 2.5.

2.7.3 Interpretation:

The bivalve-bearing association of assemblage III is broadly interpreted as the deposit of a vegetated lake margin. The micro-banded aspect of the coal seams in drill core from SH94, together with the lack of well-defined lithotypes and abundance of quasi-sapropelic shale point to the formation of peat under a relatively high-groundwater level (Hacquebard and Donaldson, 1969). The impure nature of the coal and intercalation with thinly bedded sandstone (Sd) indicates a low-lying (cf. planar) mire margin that was vulnerable to clastic incursion (probably discrete overbank flooding) due to a position proximal to a fluvial distributary channel. Flooding of the mire resulted in lacustrine conditions sufficiently oxygenated to support freshwater sharks, bottom-dwelling pelecypods and a serpulopsid worm epifauna.

Rhythmically laminated mudrock and sandstone occurs in abundance within recent lacustrine and lacustrine delta deposits of the Atchafalaya Basin (Coleman, 1966; Tye and Coleman, 1989). These lacustrine deposits also yielded polychaete worms, pelecypods, and ostracods amongst other

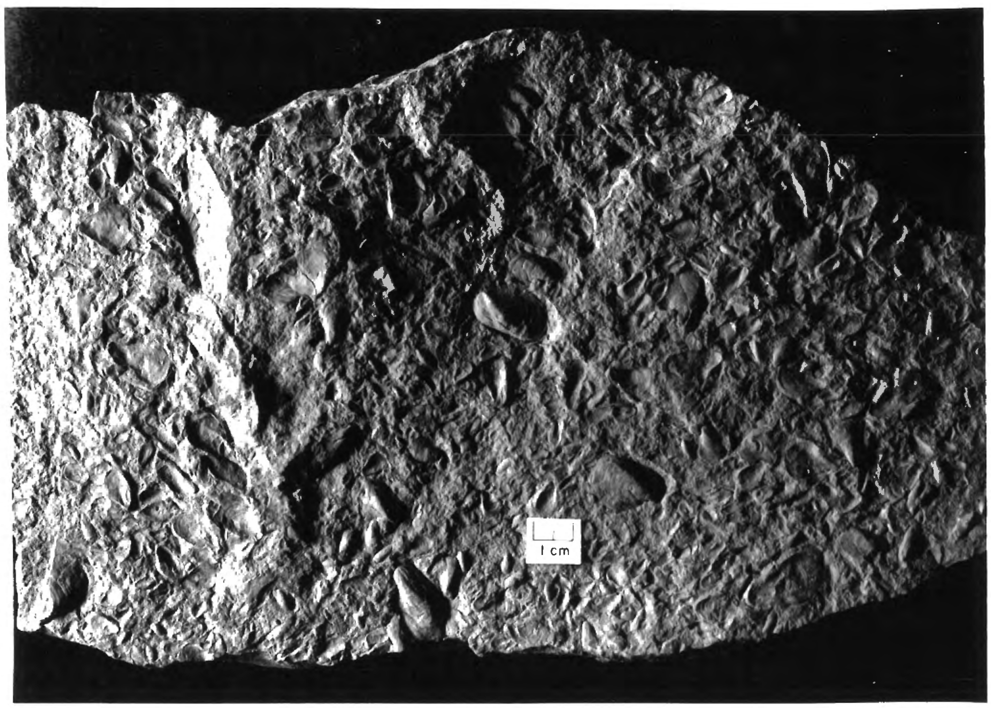


Plate 2.7 Pelecypod valves in tabular silty mudrock, exposed on Black River, Salt Springs.

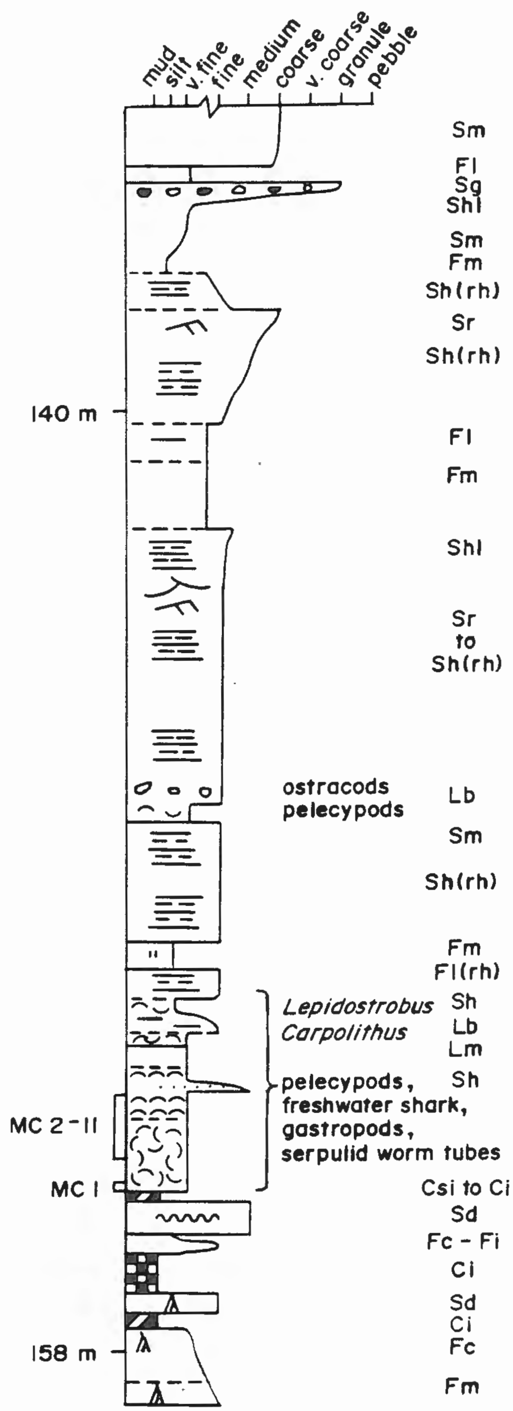


Figure 2.5 Pelecypod-bearing association of lithofacies assemblage III, corehole SH94.

invertebrate fauna. Similar stratified fine sediments containing shell fossils were given as the lacustrine facies of flood basins by Krinitzsky and Smith (1969). Coarsening-upward rhythmically stratified sandstones prevalent in the pelecypod-bearing association within drillhole SH94 are similar to the "sandy rhythmites" of Farrell (1987), from the False River area of the Mississippi Valley. These sediments were interpreted by Farrell as the deposits of a prograding natural levee.

Duff and Walton (1973) interpreted the occurrence of *Carbonita sp.* as reflecting freshwater conditions and the pelecypods "near marine" or brackish based on prevailing interpretations of salinity tolerance of the bivalves. An obvious problem arises, however, from frequent occurrence of the two together. They interpret the occurrence of lithofacies Lb overlying a coal seam as having resulted from drowning of delta-plain peats in an area marginal to the sea by a relative rise of sea-level. They refer to the influx of saline waters as an "abortive marine transgression". This contrasts with the interpretation of similar lithofacies from drillhole SH94 which were considered by Mamet (written communication, 1983) to be of definite freshwater origin. Future work on this topic potentially will have great significance for the analysis of the Cumberland Basin.

This association differs from the previously described strata of assemblage (III) primarily by the presence of the bivalve-bearing mudrock. The thinly interbedded sandstone/mudrock assemblage (III) is interpreted as a shallow interdistributary lake-fill, whereas association III B reflects deposits of the lake shore margin of low-lying peat mires. The two may therefore form a depositional continuum.

2.8 MUDROCK/MULTISTOREY SANDSTONE LITHOFACIES ASSEMBLAGE (IV)

2.8.1 Occurrence:

A comparative wealth of data exists for this lithofacies assemblage because it hosts economically significant coal seams and has been extensively drilled. The mudrock/multistorey sandstone assemblage (IV) generally occurs to the north of the poorly sorted lithofacies assemblage (II) on the southern limb and in the axial region of the Springhill Anticline. The assemblage attains a maximum known thickness of 510 m on the northern limb of the Springhill Anticline, the region which was host to the bulk of past mining.

2.8.2 Description:

Within the assemblage, three subassemblages are recognized: a) multistorey sandstone; b) coal; and c) mudrock-dominated. The relative abundance of lithofacies in this assemblage, as measured in the composite reference section from drillholes SH5, SH74 and SH99 (Figure 2.6), is summarized in Appendix A. Conglomerates (G) are absent in this compilation but occur rarely in other cores; intraformational conglomerate with mudrock clasts is included in lithofacies Se. Normally graded, pebbly sandstones (Sg) were observed as isolated scour-fills in the Rodney Sandstone within the Novaco open-pit mine but were not represented within drill core in the composite section. On average, sandstone lithofacies comprise 36% of the assemblage, and mudrock comprises 61%. Coals are most abundant in this assemblage, forming approximately 5% of the strata. Limestones, commonly exhibiting cone-in-cone structure (Lcc), account for less than 1% of the total.

The mudrock:sandstone lithofacies ratio varies significantly within the assemblage, with a general upward increase in the relative abundance of sandstone. Mudrocks are relatively abundant low in the composite section (Figure 2.6) summarized in Table 2-9; the 90 m of strata below the No. 6 seam, in drillhole SH74, have an F:S ratio of 4.3:1. This ratio decreases upward to 1.6:1 for the 134 m of strata cored in drillhole SH99 beneath the No. 2 seam. High in the reference composite section, represented by 158 m of strata above the No. 2 seam in drillhole SH5, the ratio nears unity (F:S = 1.1:1). A ratio of approximately 1:1 is typical for strata above the No. 3 seam throughout the coalfield. The inherent variability of the ratio necessitates that caution be exercised in its use as a diagnostic criterion in the definition of lithostratigraphic units.

The assemblage is cyclic in nature. Erosionally based multistorey sandstones are overlain gradationally by various mudrock lithofacies which in turn are overlain by coal seams up to 4 m in thickness. The coal is commonly abruptly overlain by mudrock which contains thin, discontinuous lenses of cone-in-cone limestone (Lcc). Horizontally stratified sandstone (Sh and Si) commonly occurs immediately beneath the erosional base of a succeeding multistorey sandstone. These cyclic sequences, from the erosional base of one multistorey sandstone through coal to the base of an overlying sandstone body, are commonly 10-30 m thick.

There are fundamental similarities between these sequences and the cyclothem as originally defined (Wanless and Weller, 1932). The major departure is the absence of a marine-influenced bed within the mudrocks which overlie coal seams of the Springhill coalfield. It is noteworthy that thin, laterally discontinuous cone-in-cone limestones (supposedly non-marine) commonly occur within the

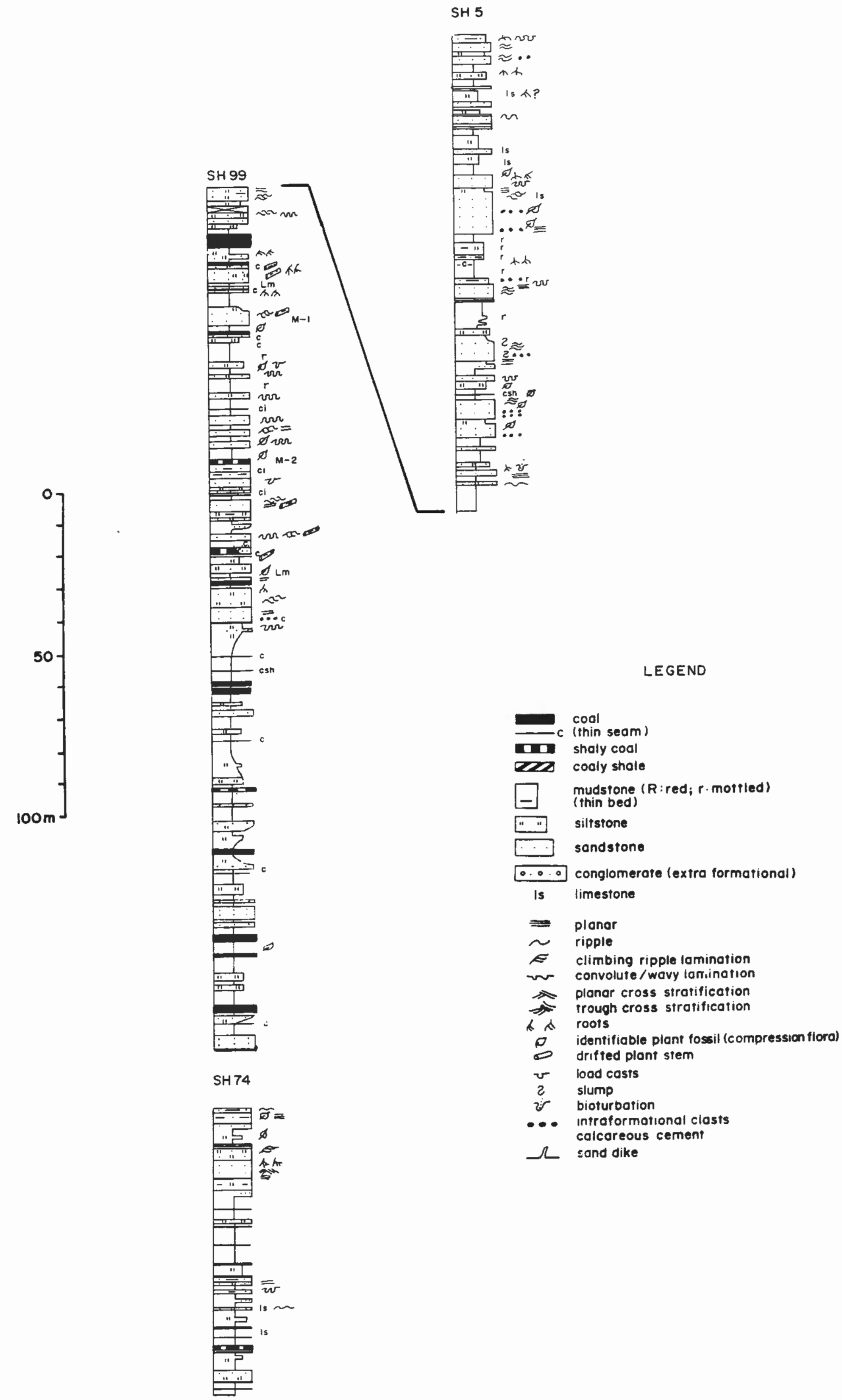


Figure 2.6 Composite section of the "coal measures", mudrock-multistorey sandstone lithofacies assemblage (IV), coreholes SH74, SH99 and SH5 (from Figure 3.1, in rear pocket).

mudrock strata overlying the coals at a similar position to the marine-influenced beds of the classic cyclothem. A similar comparison has been noted for cyclic sequences in the Westphalian D of the Sydney Basin (Masson and Rust, 1983; Bird, 1987; see Rust *et al.*, 1987 for marine vs. non-marine discussion of the Sydney strata). Cyclothems are discussed further in Chapter III and elsewhere in the thesis.

The multistorey sandstone bodies are conspicuously rare in the 200 m stratigraphic interval between the No. 2 Under and No. 7 seams (Figure 2.6) cored in drillhole SH99. Whereas this is the only recent drillhole to have cored this interval, it is not known whether the paucity of amalgamated sandstone bodies defines yet another lithofacies assemblage or whether the interval at this location (SH99) is merely an anomaly. The interval is included in the present assemblage because the lithofacies are otherwise indicative of that assemblage.

2.8.2.1 Multistorey sandstone subassemblage:

Description:

Multistorey sandstone bodies, defined as sandstone units containing amalgamated storeys separated by major scour surfaces (Pettijohn *et al.*, 1972), are one of the major diagnostic features of this assemblage. Some sandstone bodies may be stacked only in the vertical dimension (multistorey), but within the Springhill coalfield are partially amalgamated in the lateral or horizontal dimension as well (multi-lateral). The sandstone bodies range in thickness from 3 to 22 m. They are laterally persistent over distances of 1 to 4 km, based on closely spaced drillholes.

Using the descriptive terminology of Friend *et al.* (1979) pertaining to sandbody dimensions, the total multistorey sandstone bodies are described as sheet sandstones as they have a width to thickness ratio of >15:1. Stear (1983) described similarly amalgamated sheet sandstone bodies from the Permian of the Karoo Basin, South Africa, as being of composite type.

Vertical sequence of lithofacies: The sandstone bodies typically overlie gently undulating to subplanar erosional surfaces overlain successively by cross-stratified sandstone (St and/or Sp), normal-graded pebbly sandstone (Ss/Sg), thickly bedded massive sandstone (Sm) commonly containing mudstone intraclasts (Se), with no clear grain-size trend over the sequence (Plate 2.8). An ultimate fining-upward tendency from medium to fine sand is commonly evident toward the top of the vertical

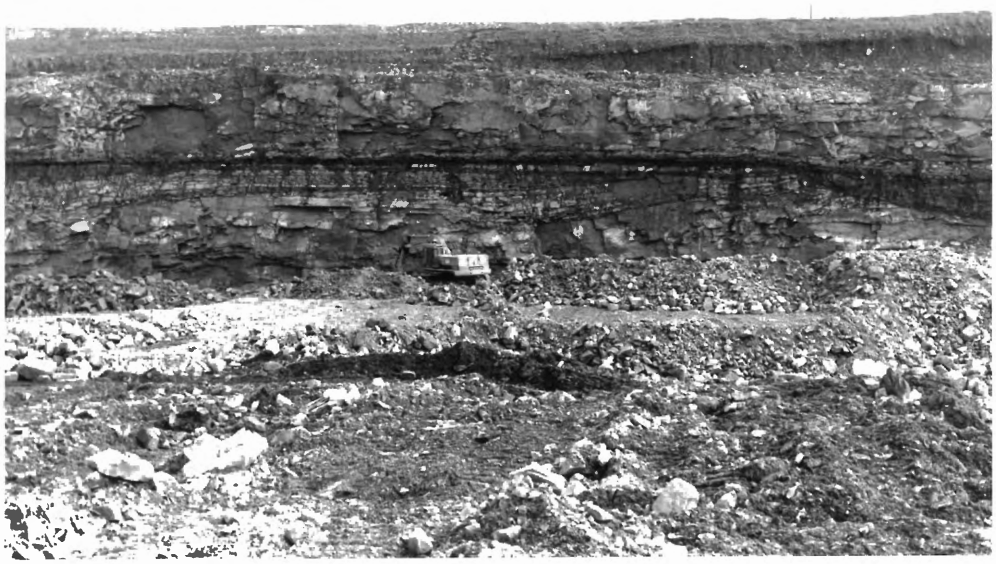


Plate 2.8 The Rodney Sandstone, exposed on highwall on Novaco open-pit mine, Springhill. Thin, 20 cm coal rests atop Rodney Sandstone. At right of photograph, edge of 6.5 m deep abandoned mud-filled hollow is evident. Horizontal to inclined heterolithic strata occur beneath thin coal seam, and sandstone-dominated complex, massive in appearance, occurs at base. Note track-mounted excavator for scale. Photograph corresponds to 20-100 m portion of Profile A-A^1, Figure 2.7.

sequence and coincides with the occurrence of thin bedded horizontal (Sh) to low-angle cross-stratified (Sl) sandstones, ripple-laminated sandstone (Sr) and sandstone with deformed lamination (Sd). In some sequences, interstratified mudrock (Fi to Si) is common within this uppermost fining upward succession.

In rare cases, fine mudrocks abruptly overlie the relatively coarse-grained cross-stratified or massive sandstones. One such example occurs at the top of the 11.5m-thick sandstone body that overlies the Rodney seam in drill hole SH55. Lithofacies St and Sp in eight cross-stratified sets, and lithofacies Sg and Sm, all of medium-grained sand, are overlain by dark grey mudrock in abrupt contact.

Internal architecture: The internal architecture of one of these sandstone bodies was obtained through detailed sedimentological mapping of the sandstone body exposed within the Novaco open-pit mine. This constitutes the only large-scale exposure of a multistorey sandstone in the Springhill coalfield. This composite sheet sandstone overlies the Rodney coal seam and is informally referred to as the Rodney Sandstone (Calder, 1984b). Mining permitted progressive mapping of the sandstone body over the period from June, 1982 to August, 1983, at various stages of its exhumation. Figure 2.7 (in rear pocket) comprises two measured sections of the Rodney Sandstone from within the Novaco open-pit. Profile A is a detailed study of the final open-pit highwall encompassing a strike length of 220 m oriented east-west, with the viewer facing south. Profile B is a smaller section, some 15 m in length, which adjoins the eastern end (0 metre point) of profile A and is oriented north-south, view to the east. This affords a limited view in the third dimension, normal to profile A. The measured sections were drawn in the field so that poorly defined forms and contacts not yet accentuated by weathering processes could be traced. Photographs and drillcore were used for verification of scale.

Within this multistorey sandstone body, a complex hierarchy of geometric forms exists (Table 2.2). The descriptive terminology employed herein is adapted from Friend (1983), Pettijohn *et al.* (1972), Allen (1983) and Gibling and Rust (1987). Multistorey units (level 1) consist of several complexes (level 2) each of which is composed of wedges, hollow fills and sheets (levels 3-5) in varying proportion.

The Rodney Sandstone (Plate 2.8) is a multistorey, sandstone-dominated but heterolithic unit approximately 15 m thick at the location of the open-pit mine. The sandstone unit is capped by horizontal beds of mudrock and fine sandstone with abundant *Stigmaria ficoides*, and an overlying coal

HIERARCHY OF SEDIMENT BODIES	LITHOLOGY	FORM	INTERNAL FORMS	THICK-NESS(M)	WIDTH (M)	SEDIMENTARY COMPOSITION	INTERPRETATION
1. Multistorey unit	sand-dominated heterolithic	composite sheet	2a,b,c,etc.	15	1000	sandstone (fine to coarse) with common mudrock inter-beds and pebbly sand(St,Sm, Sg, Fi, Si, etc.)	channel belt deposit
2. Complex (storey)	a. sand-dominated	compound sheet	4a,b,c 5	<5		cross-stratified sandstone (fine-coarse) pebbly sand-stone and minor mudrock (St,Sm,Se,Sg,Fm,Fi)	in-channel bars and dunes
	b. heterolithic	tabular/wedge hollow-fill	3,4c[1]	<5		horizontal to low-angle cross-stratified fine sand +mud(Sh,Sl,Si,Sr)	channel margin deposits
	c. mud-dominated	macroform hollow-fill		<6.5	70	sideritic mudrock with silty interlaminae; rare cone-in-cone limestone(Fm,Fi,Lcc)	major abandoned channel
3. Compound wedge/ Hollow-fill	heterolithic (IHS)	wedge with lensoid hollow-fill	4c	<3	<20 apparent	alternating mudrock/ fine sandstone in low-angle cross sets(Sl/Sh,Si, Sr)	benches and attached side (point) bars
4. Hollow-fill	a. pebbly sandstone	lensoid		<2	<10	normal graded, pebbly sandstone with intraclasts and plant stems(Sg,Sh)	single episode scour-fill from waning high-stage flow
	b. sandstone	lensoid		<2	<10	cross-stratified fine-coarse sandstone; intra-clasts(St,Sm,Se)	active channel fill
	c. heterolithic	lensoid with convex top	5[2]	<1	<10	alternating lenticular fine sandstone and mudrock drapes conforming to basal scour surface(Si)	episodic drapes deposited during repeated falling stages in semi-active channel
	d. mud-filled	lensoid to tabular					
5. Solitary Sheet/ wedge	sandstone	tabular to wedge-shape		<1	unknown	fine-med.sandstone in dm-scale planar cross-sets(Sp)	transverse bars, in-channel

1. small hollow-fills which may or may not be same as 4c
2. rarely

Table 2.2 Sedimentary bodies within the Rodney Sandstone, Springhill Coalfield.

seam approximately 0.20 m thickness (Plate 2.9). Another multistorey sandstone body overlies the Rodney Sandstone by only 2 m. The Rodney Sandstone comprises three amalgamated complexes (storeys) which are bounded by laterally extensive erosional basal contacts. Allen (1983) defined complexes as genetically related groupings of sedimentation units bounded by second-order contacts (first-order contacts being those which bound the component layers such as individual sets of trough cross-bedding). Each complex is 5 to 6 m thick, but is locally reduced in thickness through erosion at the base of a subsequent storey. Three types of complex (one example of each was evident in the Novaco open-pit) are identified primarily on the basis of lithology (see Table 2.2): A) sandstone dominated heterolithic fill; B) fine alternating heterolithic fill; and C) mudrock-dominated homolithic fill.

A. The *sandstone-dominated* complex is composed of a) hollow-fills of graded, pebbly sandstones (Plate 2.10) which overlie concave-up scour surfaces <10 m wide and <2 m deep, and b) sheets of planar cross-stratified (Sp) to large-scale trough cross-stratified (St) sandstone. Sp is fine to medium grained and occurs in sets 1 - 3 dm thick. Rarely, large-amplitude, undulating layers were barely discernible within the freshly excavated sandstone (e.g. above 20 m - mark, Figure 2.7). Alternating dm-scale beds of lenticular sandstone and mudrock which in part conform to underlying scour surfaces in the order of <10 m wide by <1 m deep are prevalent in the eastern 80 m of the measured section, particularly along a discrete horizon approximately 5 m above the base of the sandstone unit (Figure 2.7). These forms are a common but volumetrically minor constituent of the sandstone-dominated complexes.

B. The *heterolithic* complex in the measured section (Plate 2.8) occurs in the upper third of the unit, and is exposed from 0 - 90 m and again from 175-215 m of profile A (Figure 2.7). It consists of alternating 30 cm thick beds of pale grey fine grained sandstone and 5 cm thick beds of dark grey mudrock. These form hollow-fills (58-78 m, Figure 2.7) and gently dipping (5-10° apparent dip) wedges which may be bounded by erosional surfaces (0-50 m, Figure 2.7). In two dimensions parallel to hollow-fill axes, the strata may appear horizontal (Plate 2.11). Small-scale hollows are scoured locally into the inclined wedges (between 12 and 36 m, Figure 2.7). Such wedges are analogous to the inclined heterolithic strata ('IHS') of Thomas *et al.* (1987).

C. A *mudrock-dominated* complex occupies the upper third of the multi-storied unit, from 88 to approximately 190 m of the measured section in Figure 2.7. This complex is essentially a very large hollow-fill (Plate 2.12). Although the measured section depicted in Figure 2.7 appears to be a

Plate 2.9 Portion of several metre-long stigmarian axis underlying 20 cm thick clarain-rich coal seam atop the Rodney Sandstone.

Plate 2.10 Hollow infilled by poorly sorted pebbly sandstone (Sg) containing both extra-basinal and intraformational clasts and coalified stems (*Cordaites sp.*). Note abrupt grading to horizontally stratified sandstone (Sh) at top, beneath yet another erosionally based pebbly hollow-fill. Hollow-fills occur within sandstone dominated complex of Rodney Sandstone (between 0-10 m, Profile A-A^1, Figure 2.7).

Plate 2.11 Low-angle cross-stratified to horizontally stratified sandstone beds of the heterolithic complex in upper part of Rodney Sandstone. Relatively massive-appearing trough cross-stratified beds of sandstone-dominated complex at base.

transverse cross-section of the hollow-fill, it is in fact an oblique longitudinal section created by the convex, curved trace of the open-pit highwall which only just intersects the hollow-fill at its apparently thinnest edges (see Plate 2.12; also Figure 2.7)

The mudrock complex (Figure 2.8) is infilled at the base by a 40-65+ cm thick bed of sandstone which is locally ripple-laminated to planar-cross laminated with sand dikes. Above this sandstone 0.18 m of silty Fi is overlain by 1.21 m of mudrock (lithofacies Fm) containing common 2 cm thick siderite bands which are individually laterally persistent throughout the hollow-fill, and rare, very thin laminae of very fine grained sandstone. This mudrock sequence is overlain by 1.60 m of mudrock thinly interlaminated with horizontal silty sandstone (Fi); the upper 1.2 metres of this contains abundant siderite nodules. The upper 1.60 m of the hollow-fill is massive, possibly rooted, mudrock (Fm); siderite concretions are notably absent. The sequence is ultimately capped by the 0.20 m thick coal seam described earlier. Within the hollow-fill, cone-in-cone limestones (Lcc) occur as discrete lenticular bodies approximately 5 cm thick and 65 cm in diameter. These limestones occur along two separate horizons, 0.80 m and 1.79 m above the base of the hollow fill.

Sequential development of complexes:

The three complexes occur in a vertical sequence with the sandstone-dominated complex at the base, overlain in sharp (erosional to conformable) contact by the heterolithic complex. In the measured profile A (Figure 2.7), the mudrock-dominated complex overlies a deeply eroded surface which incises through the heterolithic complex and into the sandstone complex. The resulting vertical sequence (Figure 2.9) may be sandstone-heterolithic, sandstone-heterolithic-mudrock, or sandstone-mudrock, depending on the presence of and depth of erosion of the mudrock-dominated complex at any given point (Figure 2.7).

Observations made during the course of open-pit mining suggest that the "norm" is the superimposition of the heterolithic complex (horizontal to low-angle cross-stratified sandstone and alternating mudrock) over the sand-dominated complex (graded pebbly sandstone, cross-stratified sandstone, and form-concordant heterolithic hollow-fills). The mudrock-dominated complex forms a local addition to the sequence.

With respect to grain-size trends imparted by this arrangement, the sandstone-dominated complex at the base yields an irregular weathering profile with no clear trend due to the preponderance of sand with local, heterolithic pebble conglomerate and fine mudrock. The

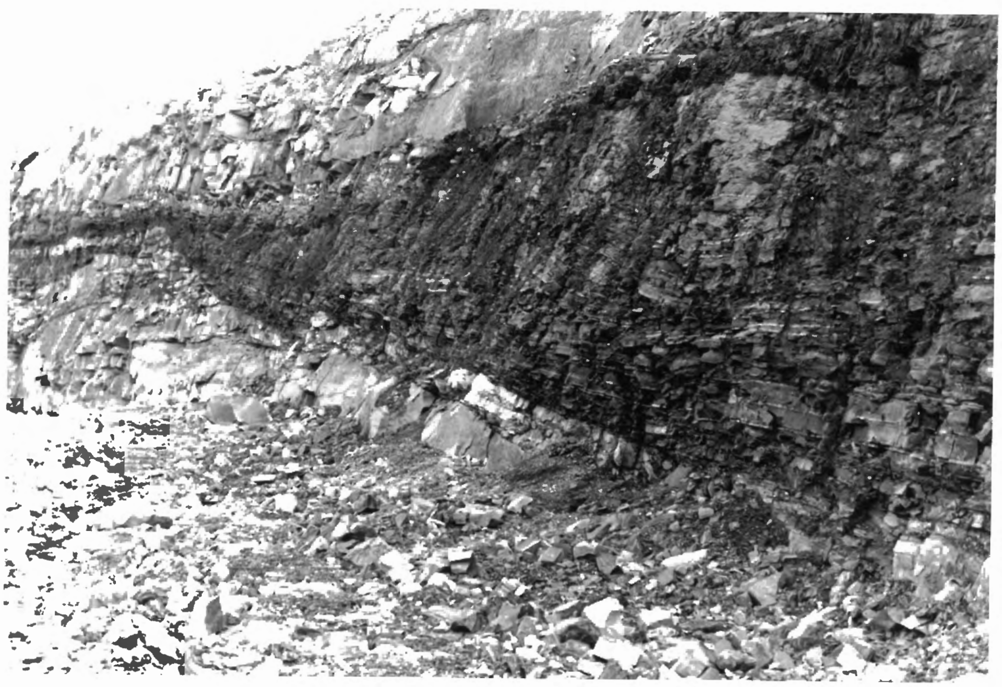


Plate 2.12 Large, abandoned mud-filled hollow incised 6.5 m into Rodney Sandstone.

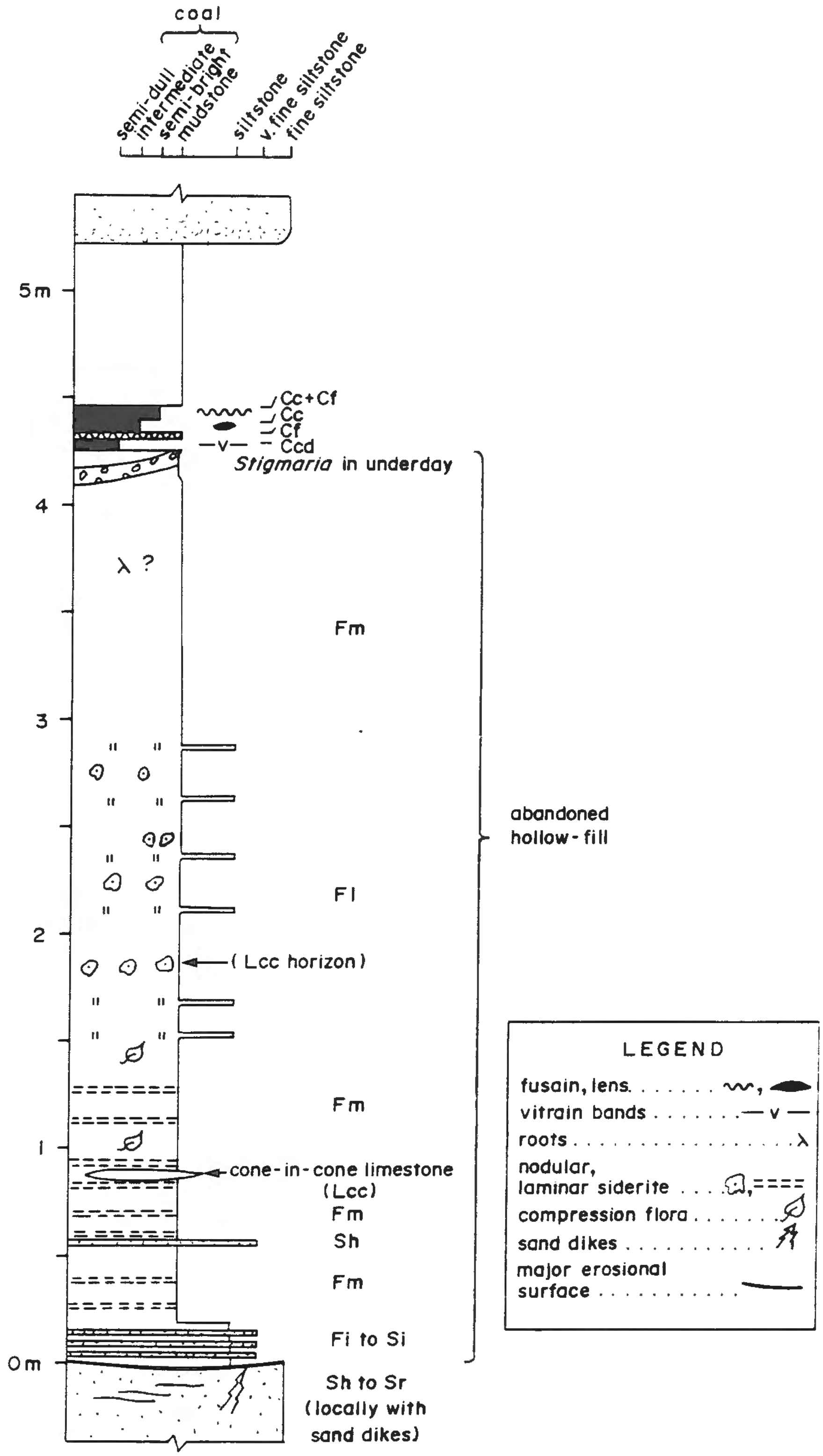


Figure 2.8 Vertical profile of mudrock hollow-fill, Rodney Sandstone (133 m, profile A-A[1]).

superimposition of the heterolithic complex with its finer grained sandstone and mudrock layers results in an ultimate fining-upward motif.

The sequential development of complexes within the measured highwall section is shown in Figure 2.9. The precise determination of the history of storying within the sandstone body involves a certain degree of interpretation. Specifically, the sequence of macro-scale erosion and subsequent infilling must be determined by tracing laterally extensive erosional boundaries. In cases where strongly contrasting lithologies are juxtaposed, this determination is relatively straightforward, but is less so in the case of juxtaposed storeys of similar lithology. Within the sandstone-dominated complex in the lower 10m of the sandstone body, it is difficult to define storeys given the relatively uniform lithology , the abundance of meso- to macro-scale hollow-fills and the non-weathered surface. A further complication is created by the orientation of some complexes (e.g. mudrock-dominated complex) parallel to measured sections.

Paleoflow indicators:

The collection of paleoflow data was limited by conditions of access to the near-vertical mine highwall. Direct measurements were permitted in places during the course of mining. The main areas for which paleoflow data were obtained are at the southeast corner of the open-pit (profile B and left of profile A); and the length of the mudrock hollow-fill (between 90 and 180m, profile A). A summary of all paleocurrent measurements from the Rodney Sandstone, including excavated portions not represented in the two profiles, is presented in Figure 2.10.

A hierarchy of sedimentary structures was determined by Allen (1966) and subsequently modified by Miall (1974). Paleocurrent indicators from the present study that were directly measured and are depicted on the profiles are: i) macro-scale hollow-fill margins (rank 3); ii) mesoscale hollow-fill margins (rank 4); iii) planar and trough cross-stratification (rank 5); and iv) primary current (parting) lineation; v) climbing, asymmetric ripple stratification and vi) fossil plant stem axes (rank 6).

Interpretation of Sedimentary Bodies and Bedforms:

The majority of the elements of the Rodney Sandstone (Table 2.2) represent deposition within a channel belt, but at different geomorphic positions and stages of flow. Hollow-fills are interpreted as infilled scours. The macroform scours (Jackson, 1975) represent major channels within

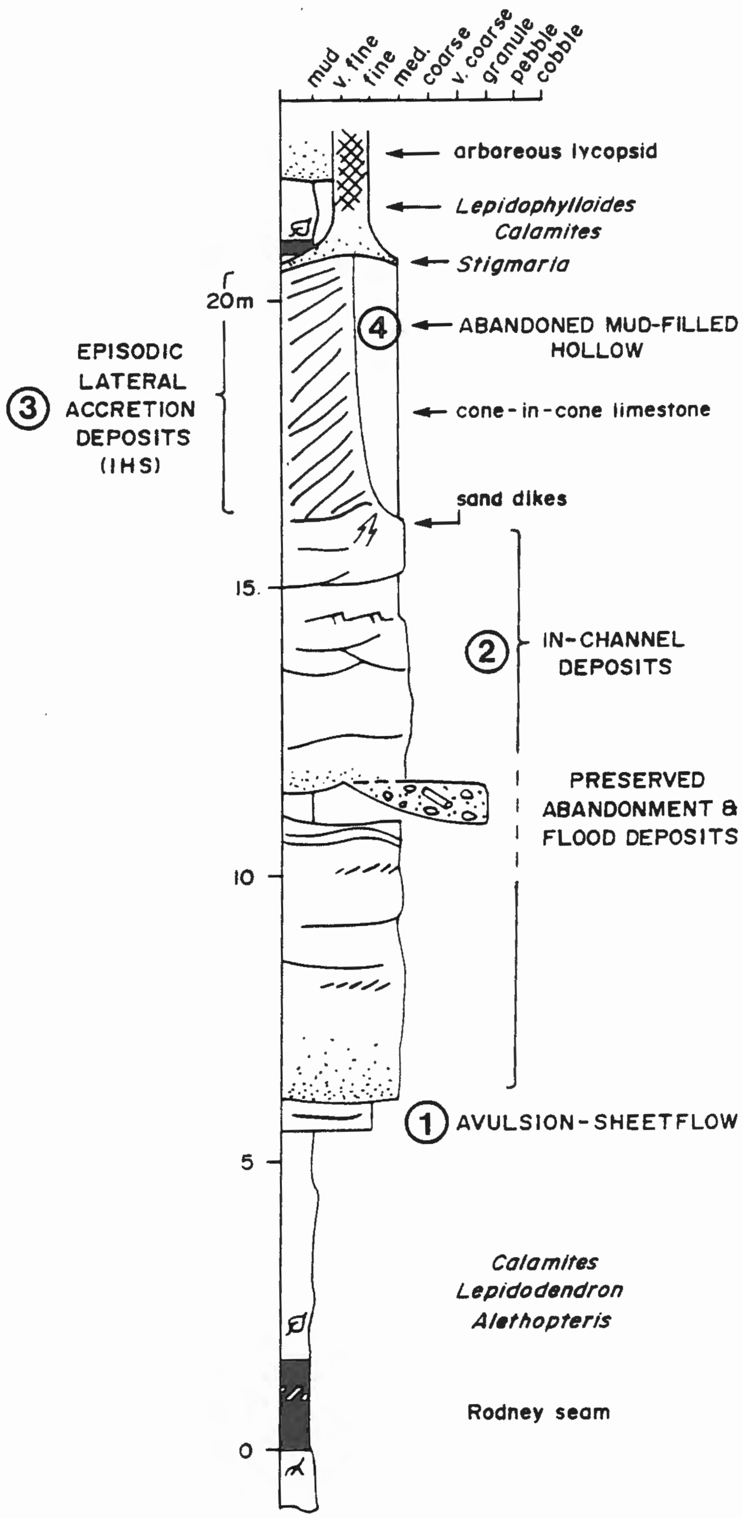


Figure 2.9 Sequential development of complexes in the Rodney Sandstone.

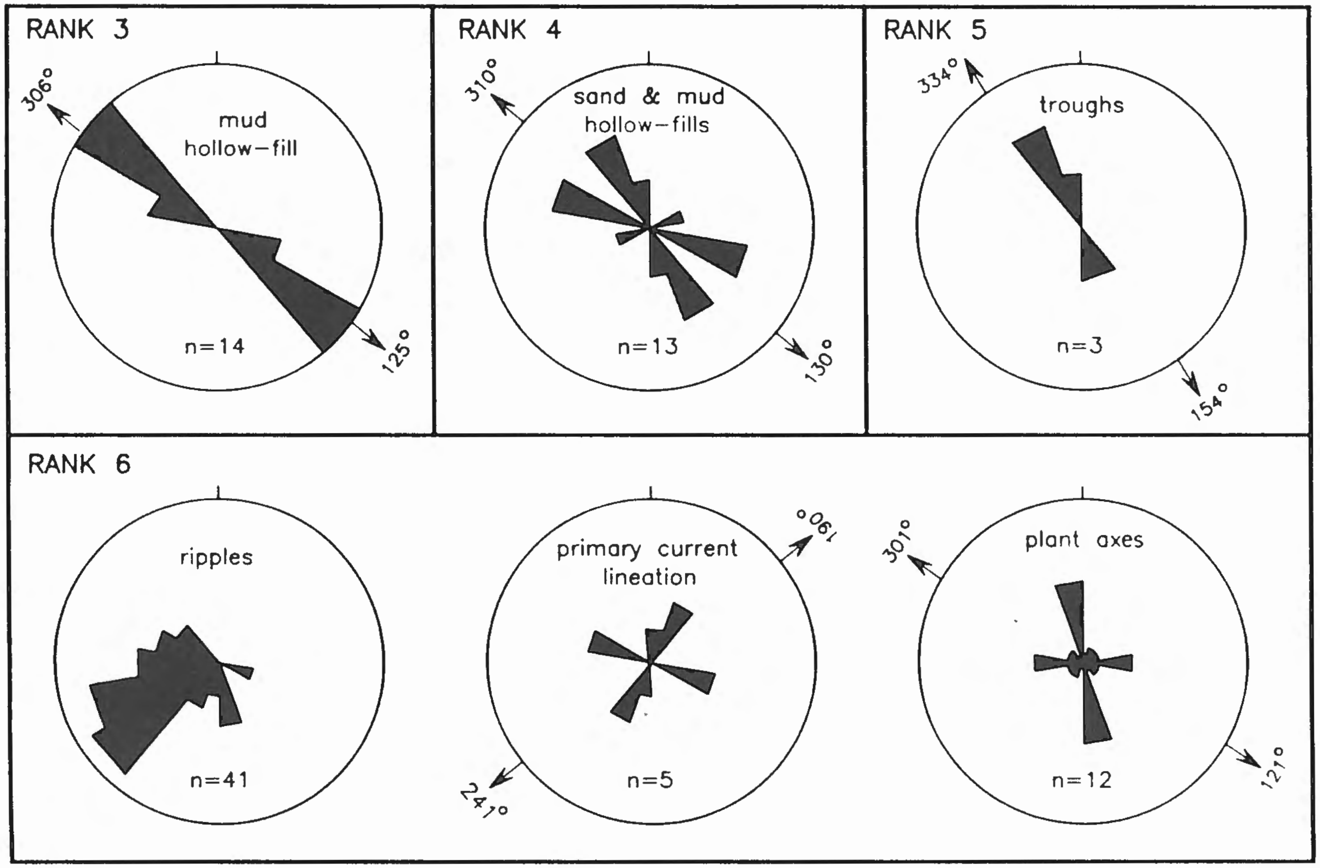


Figure 2.10 Paleoflow indicators measured from the Rodney Sandstone, ranked according to the hierarchy of Miall (1974).

the channel belt, whereas mesoform scours (Jackson, *ibid.*) represent either smaller channels or local scour-pools. The extreme variation in grain-size and texture of the infilling sediment indicates that stream competence and/or sediment grade fluctuated strongly.

Pebbly sandstone hollow-fills (4a) were probably infilled shortly after incision. The coarse nature of the pebbly sandstone suggests that bedload transport predominated. The lack of internal cross-stratification and the abrupt grading into horizontally-stratified sandstone near their tops (Plate 2.10) indicates a single event of infilling from a rapidly waning flow. The lack of modification of the basal, concave-upward erosional contact suggests that the bedload was deposited within the scour shortly after its formation, again suggesting a rapidly formed body that reflects a rapid loss of stream competence. The ubiquitous mix of intraformational pebbles, plant stems and exotic clasts indicates that flow at times eroded channel margin and floodbasin sediments, consumed riparian (river margin) vegetation and saw the introduction of immature sediment from the basin margin.

Predominantly mudstone-rich hollow-fills (4d) indicate a) that scouring and infilling were two temporally distinct events; b) that the site of accumulation of these bodies was spatially removed from areas of coarse bedload transport; or c) that a suspended load of mud characterized the river load during their formation. The large (macroscale) mudrock-dominated body (Plate 2.12) between 90 and 180 m, profile A, is interpreted as a major abandoned channel. The channel was infilled in the basal several decimeters by fine sand, locally liquefied as indicated by de-watering structures. The majority of the infilling was by mud-grade sediments deposited from suspension. Two intervals of especially quiescent conditions resulted in the formation of discrete cone-in-cone limestone lenticles (Plate 2.13; Figure 2.8). The rarity of through-flowing currents, which thus promoted development of reducing conditions, is indicated by the abundance of siderite within the mudrock fill. Therefore, although this channel was water-filled, it was isolated from active flow. Thin, cm-scale very fine sand laminae were occasionally deposited from suspension, indicating that the site of abandonment was subjected to periodic overbank flooding from (an) active channel(s). The sequence is indicative of chute and neck cut-offs in rivers of meandering planform (Walker and Cant, 1984).

Smaller-scale, mesoform scours that were infilled by mud reflect a similar dichotomy of erosion and deposition. Such bodies may reflect erosion of topographically elevated parts of the main channel at high flow stage with subsequent deposition of mud from suspension. Hollow-fills of similar dimensions (< 1 m wide by < 10 m deep) infilled by form-concordant and locally convex-upward sandstone and mudrock beds of dm-scale (4b,c) were probably formed in a similar manner. The

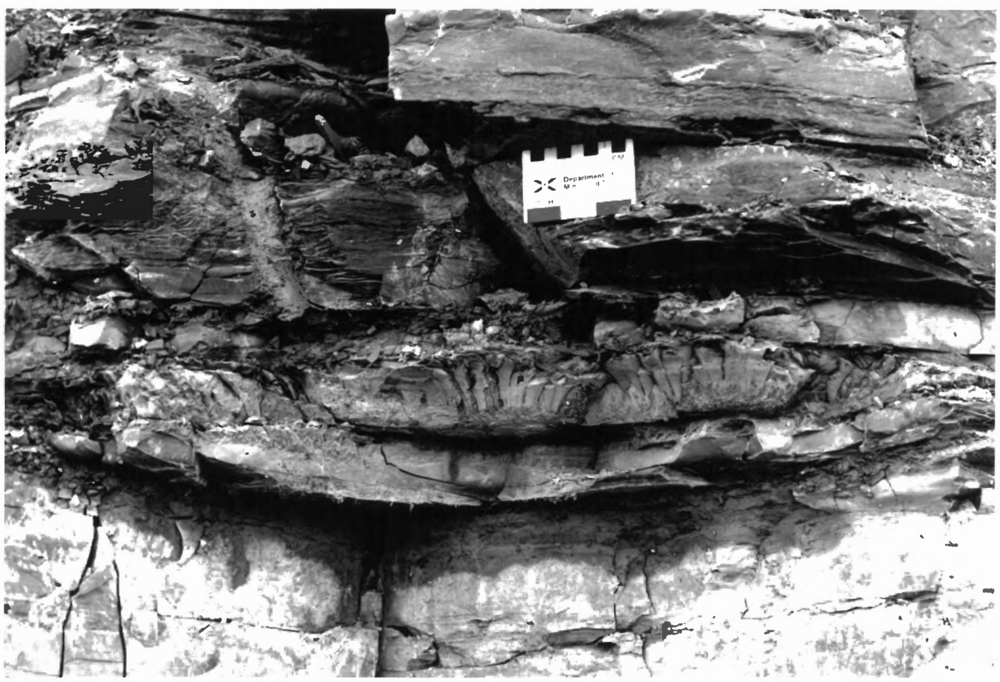


Plate 2.13 Lenticular limestone exhibiting cone-in-cone structure (Lcc) from within mud-filled hollow, Rodney Sandstone.

hollows were receptacles for episodic accumulation of sand and mud, with little or no erosion between episodes (Plate 2.14). A geomorphic setting isolated from channel-floor bedload, but perhaps topographically lower than the mudrock-filled mesoforms, is envisaged.

Sheets of large-scale (mesoform) trough cross-stratification (hierarchy level 5) indicate that fields of dunes (Harms and Fahnestock, 1965) several decimeters in height commonly occupied in-channel areas. Alternating tabular sets (each <60 cm thick) of horizontal (Sh, Sl) and planar cross-stratified (Sp) sandstone are similar to the ancient plane-bedded and cross-bedded "simple bars" of Allen (1983) described from the Brownstones (late Devonian) of the Welsh Borders, U.K. Allen interpreted these forms as having developed under upper stage flow. They probably represent broad, flat bars, locally amalgamated as composite bars. These units are common within the sandstone-dominated heterolithic complexes but were also observed from the heterolithic complex.

Hollow-fills infilled by low-angle cross-strata of sandstone, in places interstratified with mudrock (inclined heterolithic strata), are interpreted as moderate to large scale channels episodically infilled by successive low-relief bars. Stage fluctuation was not as great as for the formation of the pebbly-sandstone scour-fills. Judging from the generally concordant internal architecture of the channels, initial surges in flow were not as rigorous. Local convex upward architecture of sets of Sl may have resulted from the nucleation of a simple bar whose composite form was elongated parallel to flow; such barforms were probably rare.

Depositional History:

The depositional history of the Rodney Sandstone, represented by the sandstone body exposed in the open-pit mine, probably began following avulsion from a contiguous site. Evidence for avulsion is found in the erosional base and sharply contrasting lithology of the mudrocks overlying the Rodney coal seam and the basal sandstones of the multistorey unit (Bridge, 1984). The generally low relief of the scours at the base of the Rodney Sandstone is attributed to initial flow within multiple poorly constrained broad scours not unlike the sheetflow invoked by Farrell (1987) in her "avulsion stage" of channel belt evolution (her Figure 11b). Evidence for this multiple scour-sheetflow is found in low relief channel margins (above 145 and 188 m marks, profile A) at the otherwise planar base of the sandstone body.

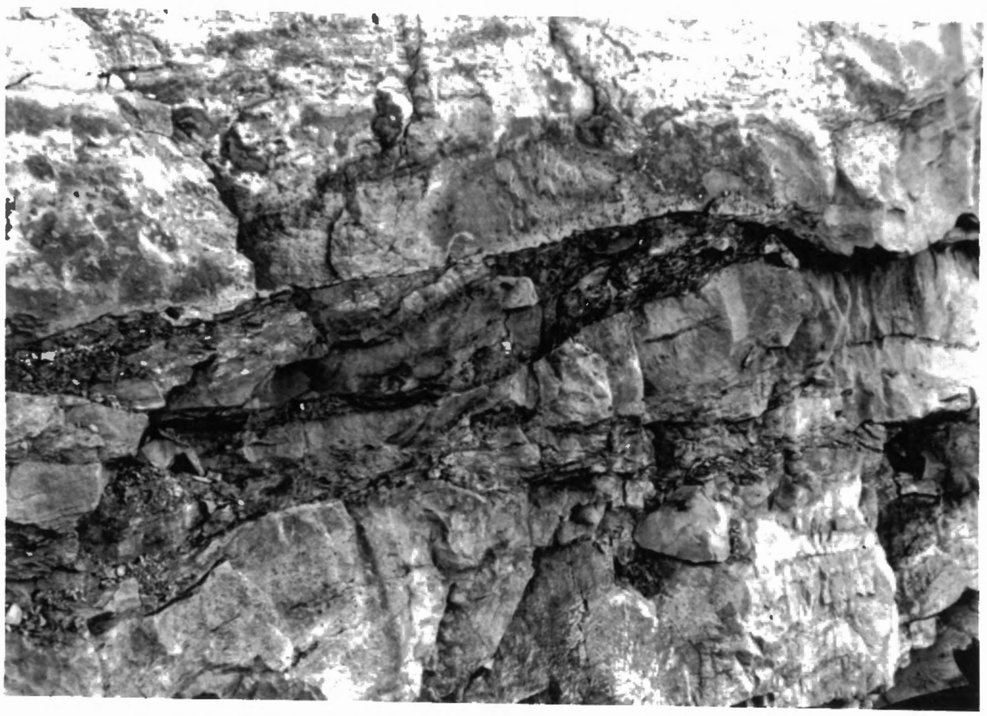


Plate 2.14 Form-concordant mudrock and sandstone beds overlying erosional surfaces in the sandstone-dominated complex of the Rodney Sandstone (above 10 m mark, Profile A-A[1], Figure 2.7).

The sandstone-dominated complex, with local, strongly heterolithic sedimentary bodies within the basal 10 m of the Rodney Sandstone, records a river system given to strong, episodic fluctuations in discharge. Because of the highly destructive events associated with the abundant scour-fills within this complex, it is difficult, if not impossible, to assess the nature of flow and resultant bedforms between peak surges.

Some insight into the nature of fluvial events that prevailed during the deposition of the sandstone-dominated complex forming the lower story of the sandstone body is provided by the examination of the bedforms exposed in profile A from 0-30 m, approximately 5 m above the base (Figure 2.7). The architecture of bedforms reveals a history of episodic lateral migration and switching of channels. Meagre evidence of periods of relatively steady lateral migration, albeit of probable short duration, is found in the easterly-dipping (to the left, in Figure 2.7) lenticular to wedge-shaped sandstone beds which may represent a laterally accreting barform that built out over an erosional channel floor. This barform can be traced downward in the direction of lateral building to undulating, in part form-concordant, fine sandstone and mudrock interbeds that overlie the erosional base. These form-concordant mounds appear to be similar to structures described in dune to plane bed transitions (Røe, 1987; Cheel, 1990). A similar origin would imply either high velocity or shallow depth of flow, or both, during their formation.

This group of form-concordant beds is incised still further to the east by a mesoscale channel that was infilled chiefly by pebbly sandstone with apparently planar stratification formed under vigorous upper stage flow. The flow was partly constrained within well-defined scours, but at times spread out as poorly constrained flow at the top of the scours. Amalgamation of pebbly sandstone-filled scours may have arisen from the switching of channels around partially emergent bars with reactivation of flow. Such periods of vigorous flow with abrupt switching of channels are thought to represent periods of increased sediment supply and flow velocity during which time the river may have taken on a lower sinuosity, braided pattern. Changes in river behaviour and pattern associated with periods of flash floods have been described from Coopers Creek, Australia (Nanson *et al.*, 1986). Braid bars in this modern arid-zone setting, however, develop in predominantly overbank areas rather than within channel.

At such flood periods during the development of the Rodney Sandstone, exotic clasts were introduced, probably by ephemeral tributaries draining the northern margin of the Cobequid hinterland. Stems of riparian flora, including *Cordaites sp.* and *Calamites sp.* (Plate 2.15) were also

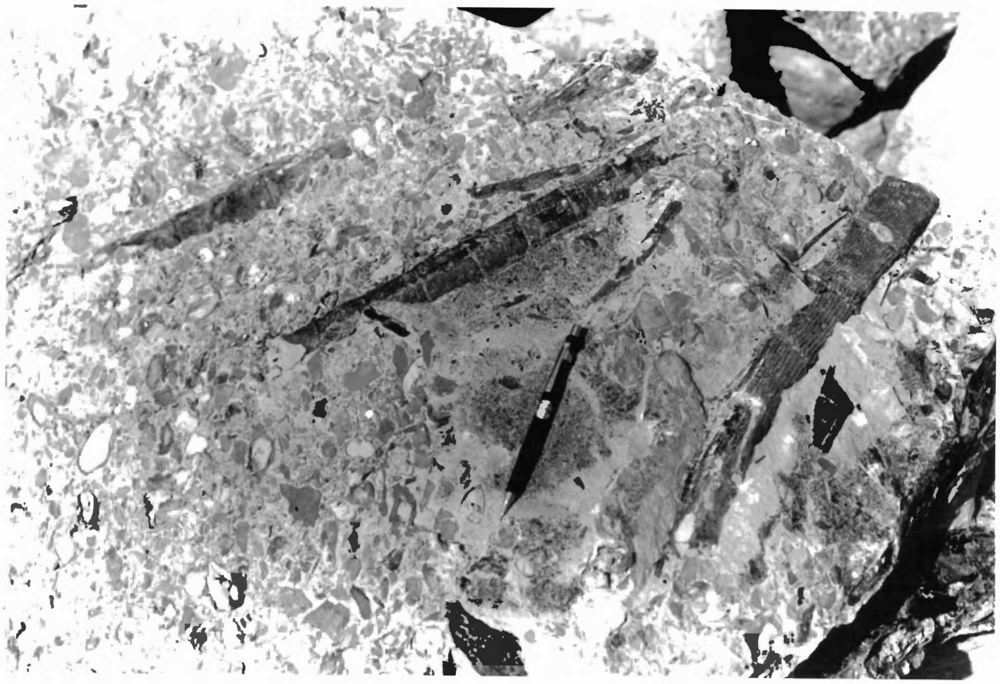


Plate 2.15 *Calamites* stems incorporated in lag deposit composed largely of siderite pebbles within sandstone, Rodney Sandstone.

swept into the waters at such times. Mudrock and siderite intraclasts were at least partly derived from erosion of intra-channel mudrock. This was unequivocally confirmed by observation of a pebbly sandstone-filled scour which had incorporated mudrock clasts. The clasts increase in size and angularity toward its erosional bank with a sideritic mudrock scour-fill from which the intraclasts were derived. The incorporation of mud clasts within channel sandstones has been attributed elsewhere to bank collapse promoted by seasonal fluctuations in the water table which causes groundwater to press toward the channel, promoting collapse of muddy banks (e.g. Coopers Creek, Nanson *et al.*, 1986) or to desiccation of floodplain muds (e.g. Cumberland Basin, Plint, 1983). In the case of the Rodney Sandstone, the intraclasts can be ascribed at least in part, to deposition and subsequent erosion by episodic flow within hollow-fills in intra-channel areas. Bank collapse may have also contributed to the formation of intraclasts, however, complete desiccation of the dark-grey carbonaceous, siderite-bearing mudrock is considered to have been unlikely. Mudrock intraclasts within the Brownstones of the Welsh Borders have been similarly interpreted as being of intra-channel origin (Allen, 1983).

The preservation of the high-stage flood deposits suggests that these scour-fills and possible emergent bars were little modified by subsequent, lower-stage flow. Rapid aggradation and lateral channel switching would have contributed to their preservation. Gupta (1983) concluded that high-magnitude flood features would be preserved if there is a periodicity (decade-scale) to the major floods, and if normal flow is insufficiently competent to modify the features.

During deposition of the heterolithic complex in the upper 5 m of the Rodney Sandstone, prevailing fluvial processes resulted in the accumulation of beds which are lithologically distinct from the lower 10 m. Within this upper storey, coarse, pebbly scours are conspicuously absent. Broad hollow-fills and shallow scours are filled by predominantly fine-grained sandstones with common interstratified mudrock, indicating that flow was either less vigorous, but nevertheless fluctuating, or that only finer grained sediment became available in the later stages of channel belt evolution.

Horizontal to low-angle cross-stratified beds (IHS) a few decimeters in thickness predominate. Channels with an approximate true width of 16 to 20 m, as deduced from combined profile and corehole data, occupied a broader channel belt. These channels, approximately 4 m deep, were infilled by low, 0.5 m thick (alternate or side?) bars. The bars apparently nucleated at times into larger convex upward in-channel barforms 1 m in thickness. Flow was deflected around and sculpted the barforms.

Such midstream bars are a common feature of modern, low-sinuosity, sand-bed rivers (Allen, 1983). They are similar, but probably finer-grained, than the symmetrical sand shoals of Allen (1983, his Figure F.) In the Brahmaputra River, similar midstream bars and larger sand flats and islands contribute to the rapid, lateral switching of shallow channels (Coleman, 1969). Diversion of flow around the bar may result in simultaneous enlargement of chute channels and eventual capture of the thalweg, leaving the previously active tract as a relict, abandoned channel (Bridge *et al.*, 1986). In certain rivers, the resulting pattern has been described as transitional between braided and meandering (McGowen and Garner, 1970).

During the final stage of channel-belt evolution a 6.5m deep strongly concave channel was incised into the broad low-angle cross-strata of the previously described complex (Figure 2.7; Plate 2.12). The channel was abruptly cut off from active flow soon after its development, as evidenced by the minimal thickness of sandstone overlying the basal erosion surface. The channel would be partly infilled by inclined, laterally accreting sandstone cross-sets, had this cut-off been related to the coincidental development of a contiguous midstream bar and related chute as in the Calamus River (Bridge *et al.*, 1986). The absence of such beds is suggestive of river avulsion or sudden channel cut-off, after which the abandoned channel was filled by permanently saturated anoxic, carbonaceous mud which was conducive to the formation of laminar siderite. Freshwater limestone (Lcc) accumulated during periods of negligible sediment input to the ponded cut-off. The whole of the abandoned channel belt was subsequently populated by a stand of arboreous lycopsids including *Lepidodendron sp.*. *Calamites sp.* may have formed an understory but may not have been precisely coeval with the lycopsids. For a substantial period of time the area was under a cover of sluggish to stagnant water which was sufficiently depleted in oxygen to permit accumulation and preservation of 1.4 m of peat represented by the 0.20 m thick rider coal seam (assuming 7:1 compaction ratio; Stach *et al.*, 1982).

In summary, the Rodney Sandstone exhibits several sedimentological features characteristic of low-sinuosity sand-bed streams. These include evidence of common channel switching (Coleman, 1969) apparently limited horizontal extent of lateral accretion deposits (Allen, 1983), and a clear tendency toward flashiness (Allen, *ibid.*). Although the subparallel orientation of the few meso- and macroscale hollow-fills (scours and channels) points to a generally low sinuosity, the NE trend of the channel body (rank 2; see Chapter III) indicates that these structures were at high angle to the overall river trend, suggesting a high degree of sinuosity. River configuration is discussed subsequently in Chapter III. The difference in sediment texture and form between the lower and upper storey may be explained by an evolutionary change in river behaviour or by an interruption of natural processes by avulsion, whereby the upper storeys were spared extensive erosive modification such as occurred in the lower storeys. The latter explanation (see also Gibling and Rust, 1987) suggests that both complexes formed part of a single, coherent deposit. Bank-attached bars, represented by the IHS, would have advanced over the sand sheet formed by the same flow system. Similar sand sheets are known to form in channels with seasonal flow, such as the Barwon of Australia (Taylor and Woodyer, 1978). The sheets develop up to the level of perennial (low stage) flow in such rivers.

Periodicity of revisitation by channel belt:

An estimate of the period between avulsion and a return of the channel belt to a site of reference (in this case profile A) can be made using tree growth and peat accumulation rates. As pointed out by Allen (1978) and by Bridge (1984), the time period represented by the vertical distance between two channel-belt sandstone bodies may include one or more avulsive episodes, therefore the time interval being scrutinized may not necessarily represent the period of avulsion.

The basis of the approximation of "return periodicity" in the Rodney Sandstone centers upon the history of the lycopsid stand that rooted atop the Rodney Sandstone (Plates 2.16, 2.17), as described previously, and was subsequently inundated by the return of the channel belt. In arriving at this estimate, it is necessary to consider a) how much time elapsed between avulsion (sudden switching of river course) and the onset of tree growth with subsequent peat accumulation; b) the length of time represented by the peat accumulation which led to the formation of the rider seam; and c) the length of time involved between the cessation of peat accumulation and the return of the channel belt to the reference site.

The growth of lycopsid trees occurred soon after channel-belt abandonment, as revealed by the soft-sediment deformation induced by root penetration within underlying sandstones. The observation that pre-existing ripple stratification in the substrate has been deflected by root appendages attached to large *Stigmaria sp.*, of lycopsid affinity, indicates that the uppermost sandstone beds were not fully lithified by that time.

The onset of accumulation of the peat overlying the *Stigmaria* horizon may have begun within several decades, but the exact time lapse is unknown. Using the range of rates of peat accumulation in modern tropical peat swamps (2.8 mm year^{-1}, Sarawak and Brunei, Anderson, 1964; 1.2-1.9 mm year^{-1}, Sarawak, Cameron *et al.*, 1989) and an estimated compaction ratio of 5:1 (see Chapter V), a

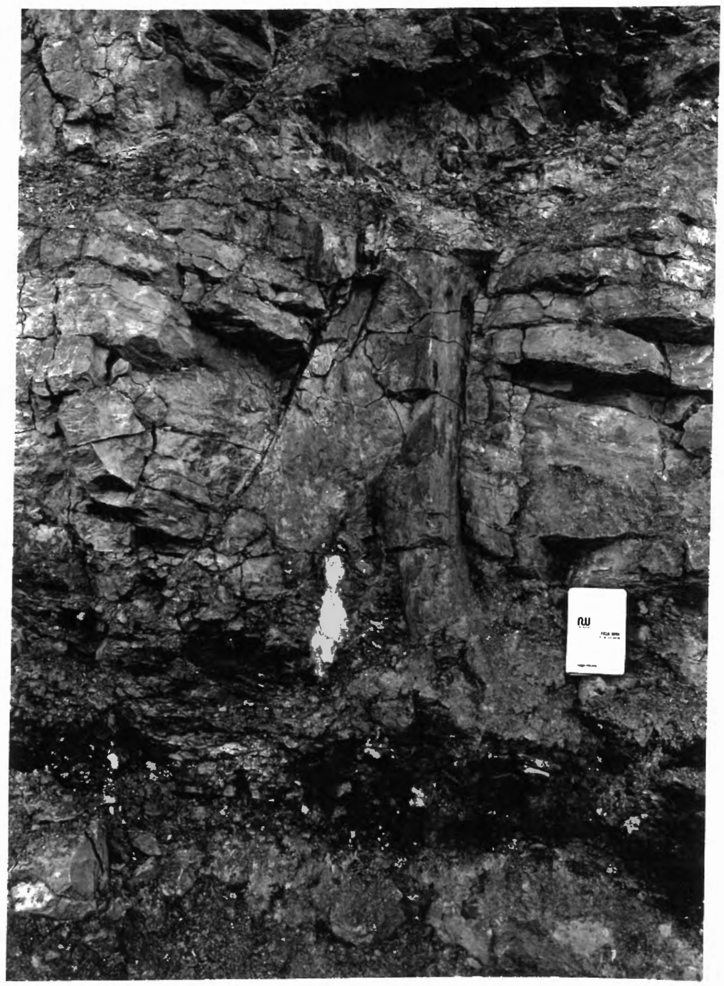


Plate 2.16 Cast of lycopsid trunk outlined by bark vitrain, within sideritic mudrock overlying thin seam at top of Rodney Sandstone.

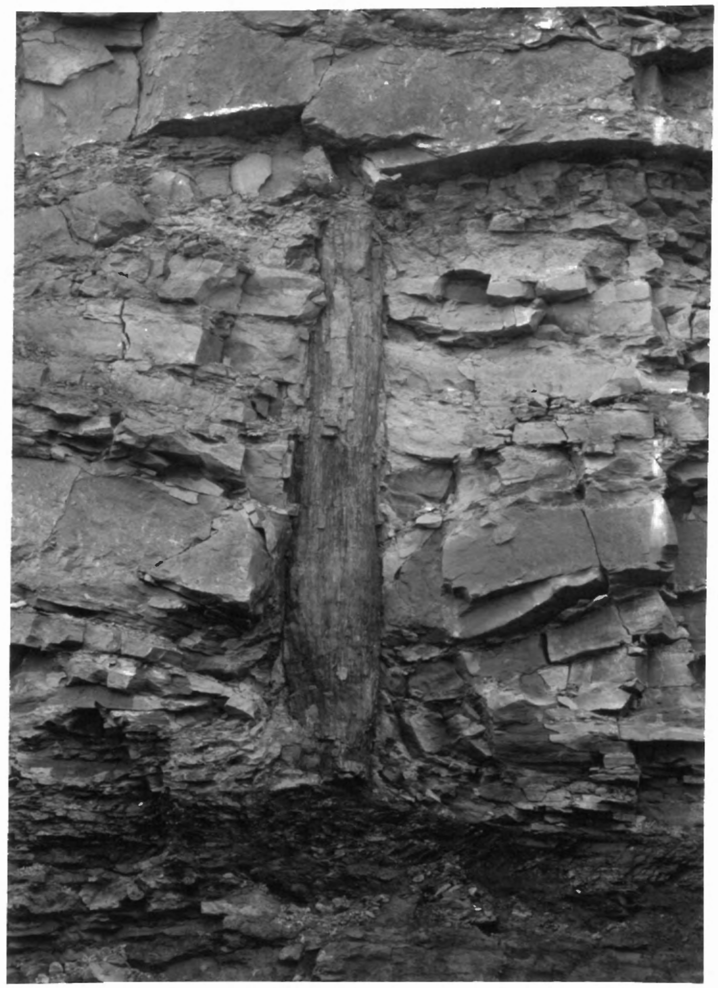


Plate 2.17 Erect lycopsid trunk, approximately 3 m in height, above 20 cm thick coal at top of Rodney Sandstone. Note strong centroclinal inclination of sandstone beds adjacent lycopsid trunk.

time period of approximately 600 years is deduced for the accumulation of the 0.20 m-thick seam.

Some measure of the time involved between the close of peat accumulation and channel-belt return can be obtained from the presence of large lycopsid trunks that were rooted within the peat and which were later partially entombed or eroded by the channel belt. Supposing that these trees began their growth at the close of peat deposition, it would not seem unreasonable to afix a century for their growth and subsequent inundation. Indeed, arboreous lycopsids may have been able to attain such dimension within decades (W. DiMichele, personal communication, 1989).

By combining these estimates, a calculation of 800 years is deduced for the minimum period of time between avulsion and re-visitation of the channel belt at this location. Whereas modern rivers tend to experience avulsion once in every 1,000 to 10,000 years within a given reach (Allen, 1978), it is likely that the estimated time period represents the period of avulsion.

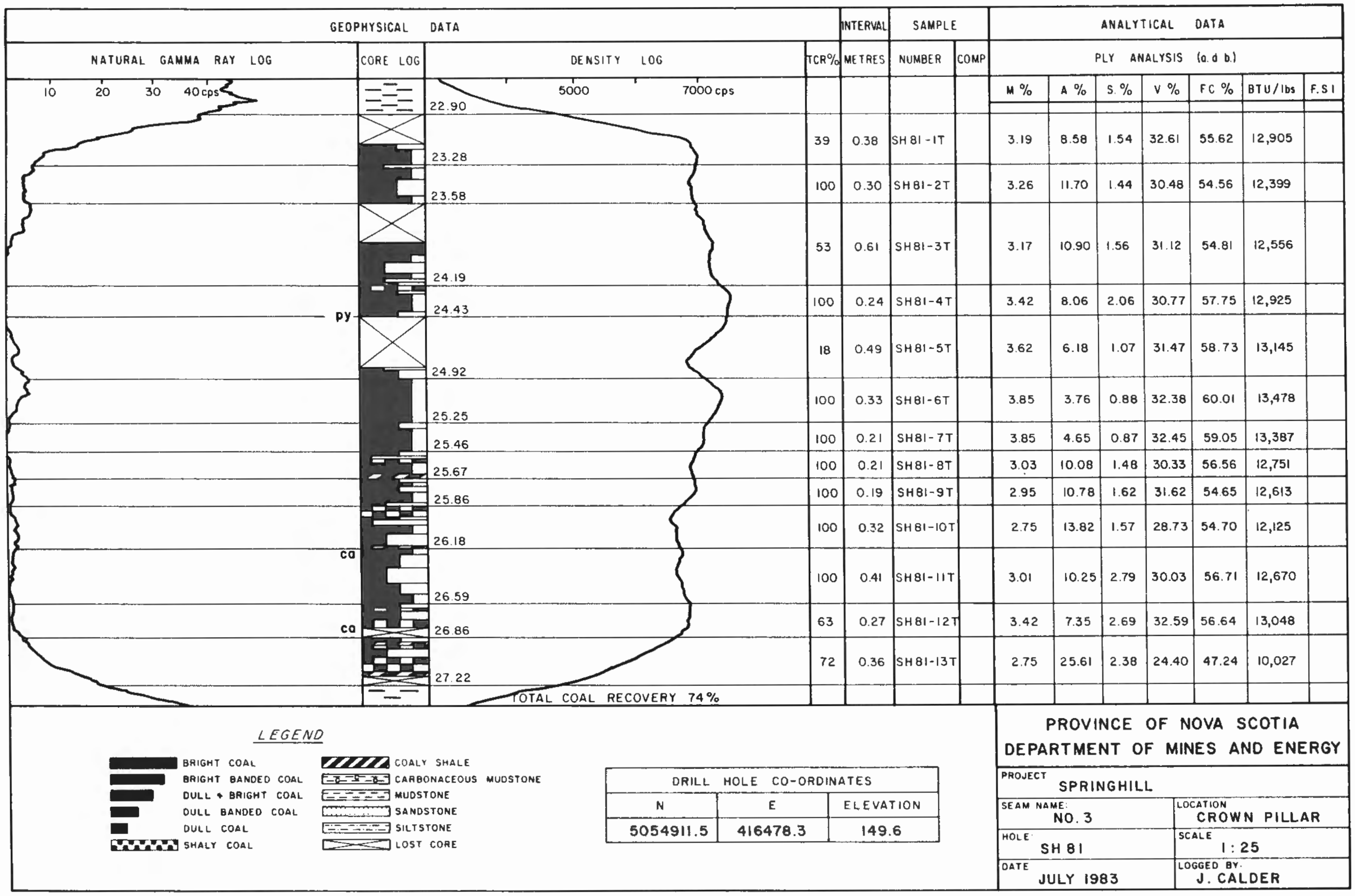
2.8.2.2 Coal subassemblage:

Coal seams within this lithofacies assemblage are typically of the banded, humic type (ICCP, 1963) and range in thickness from 5 cm to 4.3 m. Thickness of individual bands (composed of an individual lithotype) varies from less than 1 mm to 5 mm. Lithotype (i.e. coal lithofacies), ash and sulphur composition of the major coal seams of the Springhill coalfield show marked lateral variation. Clarain (embracing the writer's terms vitroclarain, clarain and duroclarain and which correspond to the bright clarain, clarain and dull clarain of MacKowsky, 1982), and impure coal (Ci) are the predominant lithotypes. Summary data of lithotype abundance for vertical sections of several major coal seams measured from drillcore are presented in Table 2.3; detailed macroscopic seam descriptions are given in Appendix B.

The intersection of the 4.32 m-thick No. 3 seam in drillhole SH81 (Figure 2.11) can be considered as a representative seam section (of the *inner zone* of mire development; Chapters III & IV). The volumetrically weighted ash (as received basis) and sulphur contents for the total seam are 10.12 and 1.65 percent respectively. Within this section, the lowest ash content obtained from a single ply sample is 3.76 percent (sample SH81-6T). This 0.33m-thick sample is composed almost entirely of clarain, with minor duroclarain, and reveals a sulphur content of 0.88 percent.

SEAM	LOCATION	LITHOTYPES (thickness- m. and percentage) Cv	Cvc	Cc	Cdc	Cd	Cf	Ci	TOTAL COAL THICKNESS
Rodney	open-pit	0.03m 1.5%	**0.97m** **49.7%**	0.27m 13.9%	0.31m 15.9%	0.19m 9.7%	0.01m 0.05%	0.15 7.7%	1.95m
McCarthy									
No. 3	SH72	tr. tr.	- -	0.19m 18.5%	0.20m 19.4%	0.22m 21.4%	0.01m 1.0%	**0.41m** **39.8%**	1.03m
	SH81	- -							3.20m
	SH27	- -	- -	- -	0.27m 12.1%	0.11m 4.9%	0.30m 13.5%	**1.55m** **69.5%**	2.23m
	WB3	tr. tr.	- -	- -	0.18 13.3%	0.22m 16.3%	0.03m 2.2%	**0.91m** **67.4%**	1.35m
	CSH79-1	- -	- -	- -	- -	0.06m 18.2%	- -	**0.27m** **81.8%**	0.33m
No. 2	SH99	0.02m 1.0%	0.29m 14.0%	0.54m 26.1%	0.53m 25.6%	0.03m 1.5%	tr. tr.	**0.63m** **30.4%**	2.07m
No. 7	SH95	0.07m 5.9%	0.12m 10.2%	**0.60m** **50.9%**	0.30m 25.4%	0.04m 3.4%	tr. tr	0.05 4.2%	1.18m
No. 6	SH95	- -	0.09m 8.1%	0.35m 31.5%	**0.51m** **46.0%**	0.09m 8.1%	- -	0.07 6.3%	1.11m
Gesner	SH74	- -	- -	- -	0.35m 41.2%	0.06m 7.1%	0.05m 5.9%	**0.41m** **48.2%**	0.85m
Sandrun	SS3	0.03m 2.3%	0.10m 7.8%	0.32 24.8%	**0.42m** **32.6%**	0.09m 7.0%	0.02m 1.6%	0.31m 24.0%	1.29m

Table 2.3 Relative abundance of coal lithotypes in various seam sections from the Springhill coalfield and Salt Springs Cumberland Basin (does not include discrete siliciclastic partings). Most abundant lithotypes are emphasised by bold type.



TCR%	INTERVAL METRES	SAMPLE NUMBER	COMP	M %	A %	S. %	V %	FC %	BTU/lbs	F.S.I
39	0.38	SH 81-1T		3.19	8.58	1.54	32.61	55.62	12,905	
100	0.30	SH 81-2T		3.26	11.70	1.44	30.48	54.56	12,399	
53	0.61	SH 81-3T		3.17	10.90	1.56	31.12	54.81	12,556	
100	0.24	SH 81-4T		3.42	8.06	2.06	30.77	57.75	12,925	
18	0.49	SH 81-5T		3.62	6.18	1.07	31.47	58.73	13,145	
100	0.33	SH 81-6T		3.85	3.76	0.88	32.38	60.01	13,478	
100	0.21	SH 81-7T		3.85	4.65	0.87	32.45	59.05	13,387	
100	0.21	SH 81-8T		3.03	10.08	1.48	30.33	56.56	12,751	
100	0.19	SH 81-9T		2.95	10.78	1.62	31.62	54.65	12,613	
100	0.32	SH 81-10T		2.75	13.82	1.57	28.73	54.70	12,125	
100	0.41	SH 81-11T		3.01	10.25	2.79	30.03	56.71	12,670	
63	0.27	SH 81-12T		3.42	7.35	2.69	32.59	56.64	13,048	
72	0.36	SH 81-13T		2.75	25.61	2.38	24.40	47.24	10,027	

Figure 2.11 Seam profile of the No. 3 seam (drillhole SH81), Springhill coalfield.

The abundance of clarain in Springhill coals is reflected in their petrographic composition, which is typically dominated by macerals of the vitrinite group. Maceral analysis of the Rodney seam (conducted by the author) and of the No. 3 and No. 7 seams (conducted by P.K. Mukhopadhyay, Atlantic Coal Institute, and now of Global Geoenergy Research) reveal vitrinite content of individual samples ranging from 23.4 to 86.9 volume percent. Liptinite group macerals range from 3.4 - 41.2 volume percent and inertinite group macerals from 2.6 - 54.3 volume percent (mineral free).

Both telinite and collinite macerals of the vitrinite maceral group are represented. Within the collinite maceral, the types telocollinite, gelocollinite, corpocollinite and desmocollinite are represented. Plate 2.18 illustrates three common modes of occurrence of vitrinite in the Springhill coals: 1) well structured with cell walls represented by telinite and cell lumens infilled either by corpocollinite or resinite; 2) poorly preserved, with variably gelified and compacted cell structure (telocollinite and gelocollinite); and 3) heterogeneous cell attritus (desmocollinite).

Coal seams of the Springhill coalfield invariably overlie rooted strata (immature paleosols) with commonly large stigmarian axes (Plate 2.9). Upright trunks of arboreous lycopsids in places protrude from seams into overlying strata (Plate 2.17). The palynoflora of the seams is overwhelmingly dominated by miospores of the genus *Lycospora* (Hacquebard and Donaldson, 1964; Dolby, 1984; 1988a; 1988b; 1991).

Ash yield and sulphur content:

Proximate and sulphur analyses reveal significant vertical variation within a given seam section, and profound lateral variation in transverse (N-S) section. So great is the lateral variability of ash yield and sulphur content that it is misleading to cite a mean value for an entire seam unless in the context of areal zones (piedmont, inner and riverine) described in the following chapter. Volumetrically weighted analyses and summaries of lithotype participation for selected vertical sections (drill core) of several of the major seams are listed in Table 2.3. Note that these are sections from the medial (inner) zones only. Ash yield and sulphur content increase greatly beyond this zone, as mentioned above.

Sulphur content of the Springhill coals is systemically variable, ranging from <2 to <5% (total seam) depending on areal position. Sulphur form analyses reveal that organic sulphur content remains relatively constant at between 0.69 and 0.81 percent, and that sulphate participation is

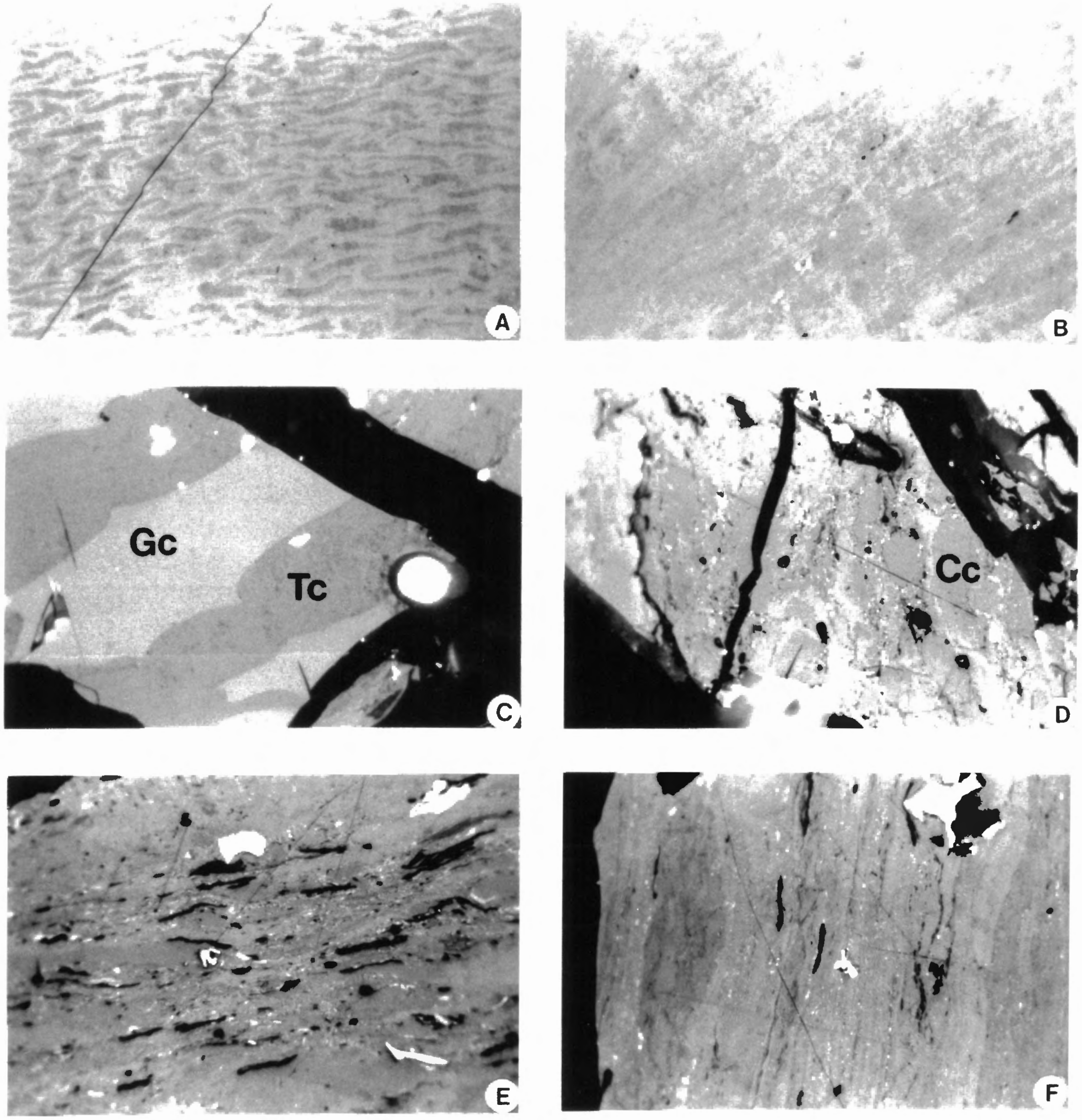


Plate 2.18 Photomicrographs from the No. 3 seam, Springhill coalfield and coal 14, MacCarrons Creek, Joggins in reflected light (x500 magnification): a) telinite (81-1T); b) telocollinite (81-1T); c) gelocollinite (Gc) and telocollinite (Tc) (MC 14); d) corpocollinite (Cc), desmocollinite and micrinite 81-7T); e) desmocollinite and sporinite (81-1T); f) desmocollinite and telocollinite (81-7T).

minimal, ranging from nil to 0.27 percent. By far the greatest variation is within the pyritic sulphur component, which ranges from 0.88 to 5.56 percent. Increases in total sulphur are invariably a result of increased pyrite. Sulphate sulphur, a weathering product of pyrite, increases coincidentally with the pyritic sulphur, whereas organic sulphur declines in inverse proportion.

The minimum ash yield for any of the ply samples from seams of the Springhill coalfield submitted to Canmet Laboratories over the period 1977-1985 is 3.35 percent. This sample (SH73-1S) comprises bright clarain and represents the upper 0.42 m of the No. 7 seam in drillhole SH73. The minimum sulphur content of any coal sample (<25 percent ash by definition) is 0.60 percent (drillhole SH9, sample #1500). The maximum recorded sulphur is 10.60 percent from the upper 0.18 m of the lower leaf, No. 3 seam (drillhole SH71, sample SH71-1LT). Sulphur declines sharply as inorganic content increases above 50 percent (coaly shale and carbonaceous mudstone).

Rank:

The coals at Springhill are predominantly of high volatile A bituminous rank with R_o max values of 0.87-0.99 for the No. 3, seam near surface. Calculated volatile matter (dry, mineral matter-free) and heating value (BTU lb^{-1}; moist, mineral matter-free) of the coals using Parr formulae fall within the designated parameters (V.M.31%; BTU lb^{-1} >14,000) set forth in the ASTM classification. Hacquebard and Donaldson (1970) determined that the rank of the Springhill coals (No. 2 seam) increases at depth to medium volatile bituminous. They attributed this trend to coalification after the basinal configuration of the strata was determined ("post-deformational coalification").

Interpretation:

Fundamentally, the coal seams of the Springhill coalfield are interpreted as fossil peat beds, which are *in situ* and essentially autochthonous. The association of erect arboreous lycopsids, well defined banding of bright lithotypes, predominance of vitrinite comprising lignin-derived telinite, telocollinite and presumably (Teichmüller, 1989), gelocollinite, resinite-filled cell lumens, and overwhelming representation of arboreous lycopsid-derived miospores, points to an origin from forested mires. The genesis of one such mire, recorded in the fossil peat of the No. 3 seam, is investigated in detail in Chapter IV.

2.8.2.3 Mudrock-dominated subassemblage

The mudrock-dominated subassemblage lies in relatively sharp contact above major (>1 m thick) coal seams, and commonly occurs beneath such seams. It usually lies in gradational contact above multistorey sandstone bodies. Within these multistorey sandstones, certain of the macroform hollow-fills are infilled by this subassemblage (Figures 2.7, 2.8 and Plate 2.12).

Mudrocks are predominantly of grey and dark grey hues, and claystone is common. Carbonaceous matter is abundant and occurs in various forms, including: microscopic carbonaceous detritus within very thin laminae (within Fcl); coarser detritus occurring as comminuted plant debris of uncertain affinity (within Fc); and well-preserved compressions of carbonized flora (within Fc, Fm). Macroflora genera found as compressions and the lithofacies with which they exhibit an affinity are given in Table 2.4. Siderite occurs abundantly as tan-coloured nodules and laterally persistent cm-thick laminae in this sub-assemblage.

Where mudrock lithofacies form the seat-earth of coal seams, rhizoconcretionary siderite nodules are common. Otherwise, evidence of rooting is rarely discernible in massive mudstone. Within mudrocks which overlie coal seams, 1cm-thick siderite laminae are ubiquitous. Cone-in-cone limestone (Lcc) commonly occurs within siderite-bearing mudrocks (Fm, Fc, Fl) in close proximity to and especially overlying a coal seam. The limestones tend to occur discontinuously along a specific horizon as discrete lenticular bodies of approximately 10 cm thickness.

The mudrock-dominated subassemblage locally includes cm-dm thick fine sandstone interstrata (Si). Internally these sandstones may exhibit horizontal stratification (Sh), ripple lamination (Sr) or soft-sediment deformation (Sd). The sandstone beds may themselves be interstratified with mudrocks (Shl).

An example of these sandstone interstrata is found in the exposure of the subassemblage above the Rodney coal seam and beneath the erosive base of the overlying Rodney Sandstone (profile A, Figure 2.7). Here, fine-grained sandstone beds, which range from 1dm to less than 1m in thickness, are laterally persistent for at least 130 m before abruptly pinching out to the east within the mudrocks. To the west, the beds thicken and locally (140-158 m, profile A) form amalgamated 2 m thick, erosive-based cosets inclined at approximately 15 degrees to the west.

FLORA	NO. OF RECORDS	LITHOFACIES AFFINITY (decreasing abundance)
Sphenopsids		
Calamites sp.	106	Fm, Fc, Fl, Sm, Fi, Se Shl, Ci, Ch, Sh, Fcl, Fg
Asterophyllites	2	Fm, Ci
Gymnosperms		
Cordaites sp.	41	Sh, Fm, Fc, Sm, Sg, Ci, Fl, Fi, Sd
Arboreous lycopsids		
Sigillaria sp.	7	Ci, Fc, Ch, Sh
Lepidodendron sp.	3	Fl, Ci, Fm
lycopsid indet.	5	Sh, Ci, Sd, Fm
Lepidophylloides	4	Fc, Fm
Lepidostrobus	2	Sd, Sh
Stigmaria	16	Fm, Sm, Sh, Sd, Fcl, Fl
Pteridosperms		
Alethopteris sp.	1	Fm
Carpolithus	1	Sh

Table 2.4 Fossil flora recorded by the author from the Springhill coalfield, and their lithofacies affinities.

Interpretation:

The mudrock-dominated subassemblage is interpreted as the deposit of a flood basin adjacent to a major channel belt(s). This floodbasin was much of the time a recipient of only fine suspended sediment. The well-preserved, carbonized compression flora and abundant siderite bespeak water-saturated muds with reducing conditions. Similar siderite-bearing carbonaceous muds have been described from regions of the Atchafalaya river basin where standing water is present perennially due to poor drainage (Coleman, 1966). Such deposits have also been described from the contiguous False River area of the Mississippi (Farrell, 1987). Both authors describe the depositional site as a poorly drained swamp. (Coleman used "swamp" as a wetland term, apparently without consideration of peat formation, in the manner of Penfound, 1952).

The nature of plant preservation and the occurrence and nature of siderite within these poorly drained wetland sediments shed further light on conditions within and above these accumulating muds. The partial preservation of the original organic matter of the entombed flora can be attributed to toxicity arising from humic acid derivatives which became concentrated in the sediments and waters (Twenhofel, 1961; Coleman 1966). The survival of delicate plant structures (e.g. attached *Alethopteris sp.* frond, Plate 2.19) necessitates minimal transport and low energy levels which verifies sluggishly moving to stagnant water (Scott and Collinson, 1983).

Siderite is thought to have formed at a very early diagenetic stage within permanently saturated soil (Wilson, 1965; Reading, 1978; Ho and Coleman, 1969). The precipitation of iron carbonate at organic-rich horizons within anoxic muds (Twenhofel, 1961, referenced in Coleman, 1966) may partially explain the occurrence of laminar siderite within the plant-bearing mudstones. Biochemical processes, specifically the depletion of CO_2 in the substrate through photosynthesis (Fairbridge, 1967), may have promoted the formation of rhizoconcretionary siderite. Such tubular siderite nodules, nucleated about a carbonaceous rootlet, commonly occur beneath coal seams of this association. Similar siderite nodules have also been reported from underclays of the South Wales Coal Measures (Wilson, 1965).

Interbeds of gradationally and locally erosionally-based sandstone (Sh, Si, Sr, Sd) are interpreted as proximal floodbasin deposits which formed largely from falling-stage sheetflood. Such

Plate 2.19 Intact frond of *Alethopteris sp*. within siderite-bearing claystone overlying the Rodney seam.

deposits were theoretically differentiated into levee and crevasse splay deposits by Elliot (1974), however in his study of the Brahmaputra river, Coleman (1969) determined that the natural levees largely comprise crevasse splay deposits. The subdivision of "levee" and crevasse splay in the ancient record may therefore be impractical.

The internal geometry of these proximal sediments, described by Coleman (1969) as interfingering and overlapping lenses of fine sand and mud, closely resembles the locally amalgamated lenses of sandstone earlier described overlying the Rodney seam in the west half of profile A, Figure 2.7a. Discrete, erosional-based scour-fills elsewhere within the mudrock-dominated subassemblage may represent crevasse channels proximal to major channel systems (Coleman, 1969).

Ripple-laminated sandstone (Sr) capped by mudrock is prevalent within the crevasse splay deposits of the Brahmaputra. These represent deposition from an episode of waning flood (Coleman, 1969). Simple fining-upward cycles as these were termed "rhythmites" by Farrell (1987). Such deposits may be rapidly colonized by plants, resulting in extensive bioturbation. Lithofacies Si of the mudrock subassemblage is analogous to these thin, multiple rhythmites.

It has been established that proximal levee/splay deposits and poorly-drained substrates of peat-forming sites were vegetated as evident from rooting. The general paucity of preserved roots within the laminar siderite-bearing and massive mudrocks, however, raises the question, "Were the perennially saturated/submerged floodbasin muds vegetated?" Coleman (1969) felt that the floral remains within his similar "poorly drained swamp" deposits represented an essentially *in situ* flora, but it is conceivable that the little-transported compression flora from the subassemblage of this study were derived from proximal areas marginal to the river(s) (cf. Schiehing and Pfefferkorn, 1984). Alternatively, the massive aspect of the mudrocks may arise from root-bioturbation and collapse (cf. Coleman, 1969; Farrell, 1987) and the apparent lack of root fossils may be a preservation phenomenon.

In summary, the mudrock sub-assemblage is interpreted as the deposits of perennially saturated or submerged low-lying wetlands (cf. backswamps of Farrell, 1987; beels of Coleman, 1969) and marginal levees (cf. levees and crevasse splays of Farrell, 1987 and natural levees of Coleman, 1969). Together these comprise floodbasin deposits (Allen, 1965) with the muddy wetland generally situated in a protected, distal setting and the levee/splays in a relatively proximal area adjacent to a major river channel belt.

2.9 VARIEGATED MUDROCK/MULTISTOREY SANDSTONE ASSEMBLAGE V

2.9.1 Occurrence:

The variegated lithofacies assemblage (VI) is best exposed on Harrison (Mountain) Brook south of the Athol road in the east of the coalfield (Figure 1.3). Within existing drill core, the assemblage is best represented in drill hole SH5. Mudrocks are poorly exposed, hence under-represented on Harrison Brook. Drill core from SH5 provides a more complete stratigraphic record, although mudrocks in the drillcore are in places significantly weathered due to exposure to the elements for a considerable period subsequent to drilling.

2.9.2 Description:

As its name implies, this lithofacies assemblage is defined in large part by variegated mudrocks, with abundant reddish and mottled shales. The assemblage is further identified by a concomitant decline in the abundance of coal seams. It can be considered a lithological transition between the coal-bearing and red mudstone assemblages IV and VI. The relative abundance of lithofacies in the assemblage as it was intersected from 143.90-312.60 m in drillhole SH5 (Figure 2.12) is given in Appendix A. In this section, the ratio of sandstone to mudrock is 1:1. Nonfossiliferous (at macroscopic scale) limestones are rare, and conglomerate lithofacies absent. Although coals are not present in this section, they occur rarely elsewhere in this assemblage, particularly in the vicinity of the present Springhill Anticline. Coal seams of this assemblage commonly exhibit an ash content elevated over that of the coals of the previously described "coal measures" (assemblage IV).

Nodules are common within mudrocks, in particular lithofacies Fm. The form of concretion varies significantly with colour of the host lithofacies (Table 2.5). Siderite concretions favour grey mudrocks, whereas calcite nodules occur preferentially in most reddened lithofacies. Similar relationships have been observed within the late Westphalian Morien Group strata of Nova Scotia (Rust *et al.* 1987). Solitary silty limestone beds occur rarely within predominantly reddish-brown mudrocks containing calcite nodules (e.g. SH5, 233.77 m and 265.70 m). At 265.70 m in drillhole SH5, the calcareous zone comprising Lm and nodules is rooted.

Multistorey sandstone units are < 12.84 m thick and appear to resemble those in the underlying (grey) assemblage IV, however paucity of large-scale outcrop in this assemblage precludes

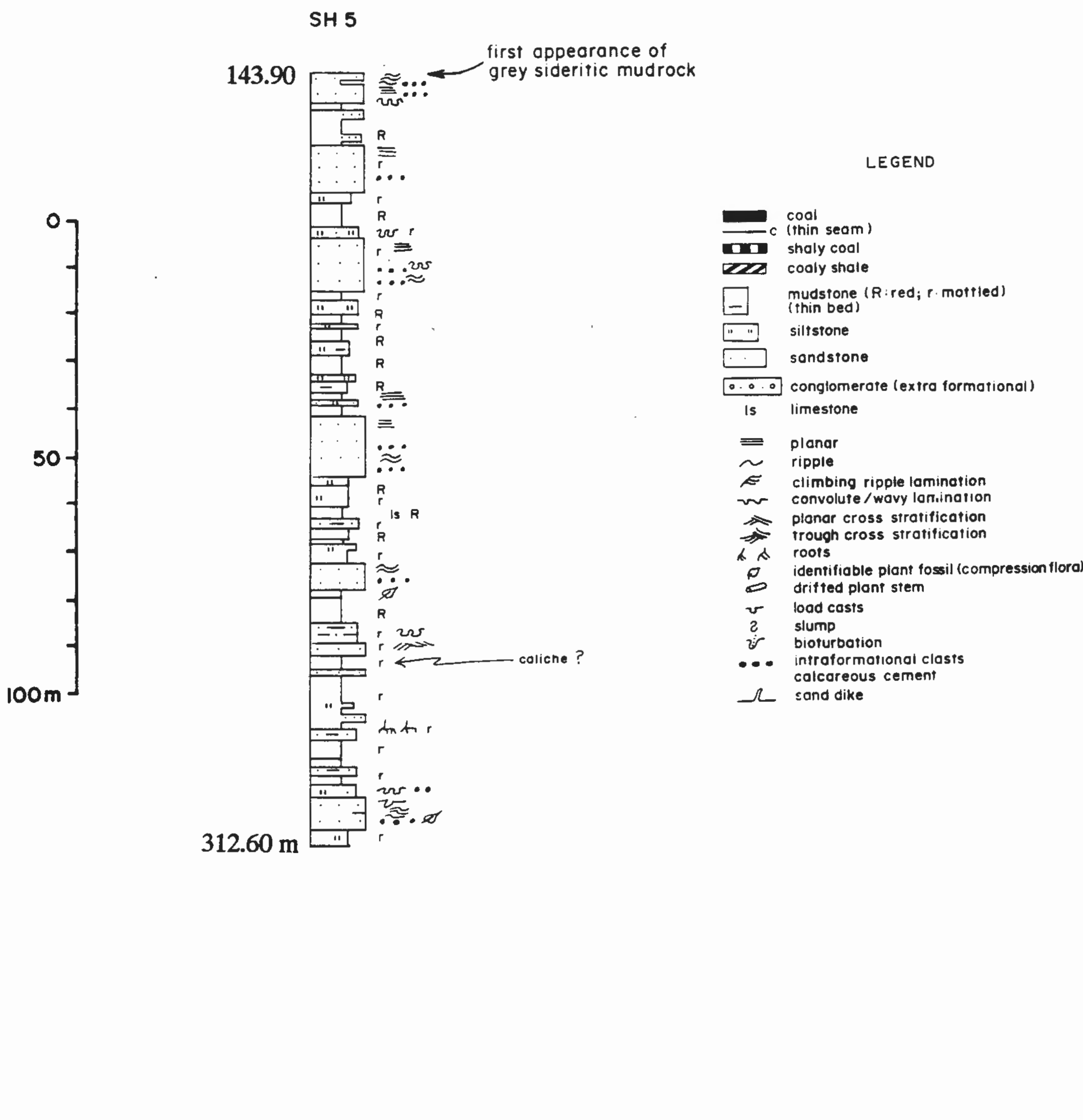


Figure 2.12 Measured section of lithofacies assemblage VI, 143.90-312.60 m, drillhole SH5 (from Figure 3.1, in rear pocket).

HOST LITHOLOGY		CONCRETION TYPE		
Colour	**Lithofacies**	**Colour**	**Apparent Composition***	**Form**
Grey to dk.grey; rarely mottled	Fm (to Fc)	tan	siderite	laminae, lenses & nodules
grey; green-grey; mottled	Fm; also Fi-Fl	pale tan to brown	calcite/ siderite	nodules; rare laminae
mottled; red-brown to red-grey	Fm	red-brown	calcareous	nodules
red-brown; rarely red-grey	Fi; also Fm (and P?)	yellow-white to pale grey	calcite	nodules

* based on macroscopic determination only.

Table 2.5 Lithofacies - concretion affinities within variegated mudrock/multistorey sandstone lithofacies assemblage V; drillhole SH5, 143.90-312.60 m.

detailed comparison of meso- to macro-scale bedforms and internal architecture.

2.9.3 Interpretation:

The environments of deposition are interpreted similarly to those for the underlying assemblage IV, the so-called "coal measures". The paucity of coal and the abundant reddening of mudrocks, however, point to a periodically lower groundwater table in this variegated assemblage, hence a well-drained floodplain. The fact that reddish and mottled mudrocks are interstratified with, and in places laterally transitional to, grey mudrocks together with the non-reddening of multistorey sandstone, suggests that local environmental conditions resulted in early diagenetic reddening. This relationship does not support reddening from widespread climatic change (e.g. increasing aridity) or from regional, wholesale diagenetic oxidation as proposed for the younger strata of the Pictou Group in the Tatamagouche syncline (Ryan, 1986).

The presence of carbonaceous matter (rootlets?) within the greenish-grey portion of mottled mudstones is evidence of gleying (Robinson, 1949). This process occurs where reducing and oxidizing conditions alternate, and is enhanced by the presence of organic matter and neutral to high pH. Gleying is related to fluctuating water levels whereby the soil is alternately wetted and dried (Bown and Kraus, 1981; Robinson, 1949). The presence of gleysols and associated calcite concretions ("glaebules" of Brewer, 1964) indicate that soil-forming processes related to fluctuating groundwater levels were important during the deposition of this assemblage. Common rooting and a general lack of preserved macroflora in the lithofacies assemblage indicates that plants flourished, but that aerial plant parts were not preserved under conditions of elevated Eh. The origin of nodules, both siderite and calcite-bearing, is at least in part attributed to rhizoconcretionary processes in the B soil zone. The calcite rhizoconcretions may be incipient/early calcrete.

2.10 RED MUDROCK/LITHIC SANDSTONE LITHOFACIES ASSEMBLAGE VI

2.10.1 Occurrence:

The assemblage outcrops west of the town of Springhill along the western border of the map area. Precise knowledge of the basal boundary is hampered by lack of outcrop. The assemblage occurs in the upper 143.90 m of drillhole SH5.

2.10.2 Description:

The red mudrock/lithic sandstone assemblage exhibits only subtle lithologic differences from the previously described variegated assemblage. Chief among these differences are: 1) virtual absence of greyish mudrock; 2) rarity of siderite concretions; 3) no intercalation with coal-bearing strata; and, 4) apparent compositional change in multistorey sandstones from quartzose and/or feldspathic types in underlying lithofacies assemblages to lithic type (litharenite; terminology of Folk, 1968), based on macroscopic observation. Confirmation awaits thin section analysis. Relative abundance of lithofacies in the red mudrock/lithic sandstone assemblage from 3.53 - 143.90 m, drillhole SH5 (Figure 2.13) , is given in Appendix A.

Dark grey mudrock (Fm) is rare in drillhole SH5. It occurs only in close spatial association with multistorey sandstones, either as channel fill within the sandstone unit or directly beneath the erosional base of such a unit (Figure 2.14). Vertical profiles of multistorey sandstones from measured sections of core and outcrop are presented in Figure 2.14. Outcrop sections are from Harrison Brook in the southwest of the map area, 250 m upstream from where the stream intersects the abandoned Cumberland Railway and Coal Co. rail bed.

The multistorey sandstone units cored in drillhole SH5 represent the thickest measured within the strata of the Cumberland Group in the southern Cumberland Basin. The unit depicted in Figure 2.14a is 34 m in thickness, and is typified by a very high percentage of crudely stratified immature (weathered?) grains; low angle to horizontal stratification predominates. At two locations, 0.15 and 0.58 m-thick dark grey mudstone (Fm) abruptly overlies lithofacies Se. The upper mudstone bed at 28.93 m hosts sideritic laminae which are rarely found in this assemblage. The mudstone overlies a rooted sandstone containing scattered mudclasts. In direct contact with the top of the Fm unit is a 0.25 m thick sooty brown limestone with contorted vitrain laminae (Lcw). A second occurrence of this limestone was observed at 43.93 m within Sh, overlying Se and underlying a third, massive grey mudrock unit.

In outcrop on Harrison Brook, the multistorey sandstones contain fewer immature lithic grains and mud clasts, although the latter are locally in high concentration (e.g. bottom of measured section, Figure 2.14b). Sets of shallow trough cross-stratified sandstone (St), 10-30cm in thickness, and medium-bedded Sh comprise the bulk of the sandstone bodies.

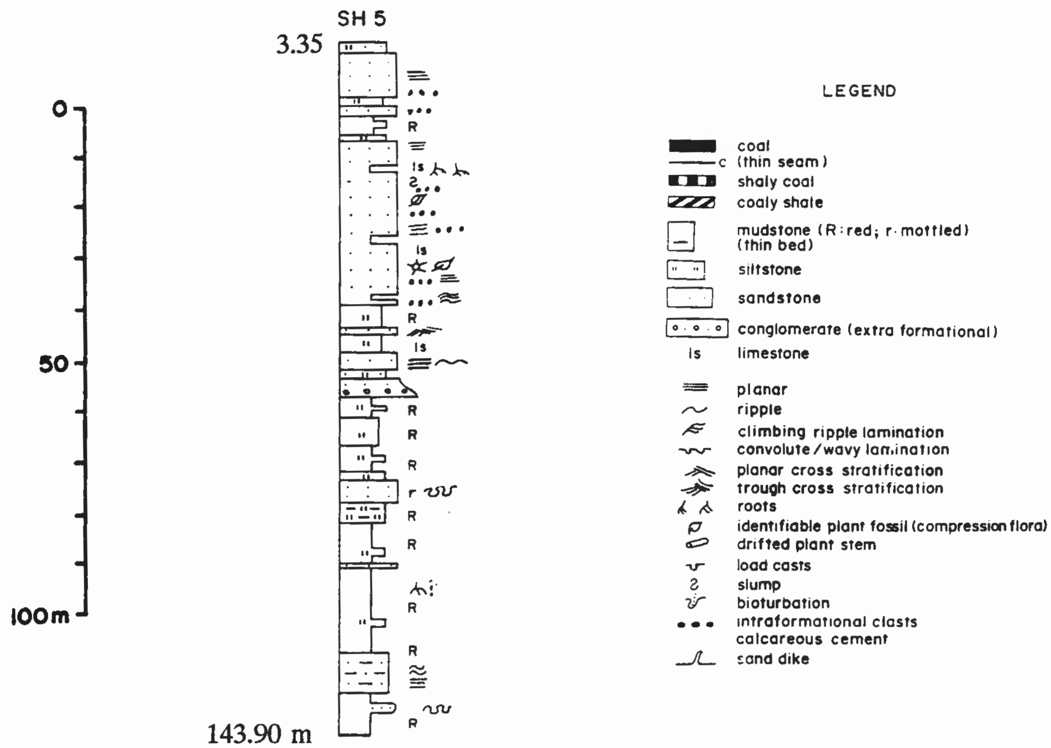


Figure 2.13 Measured section of lithofacies assemblage VI, 3.35-143.90 m, drillhole SH5 (from Figure 3.1, in rear pocket).

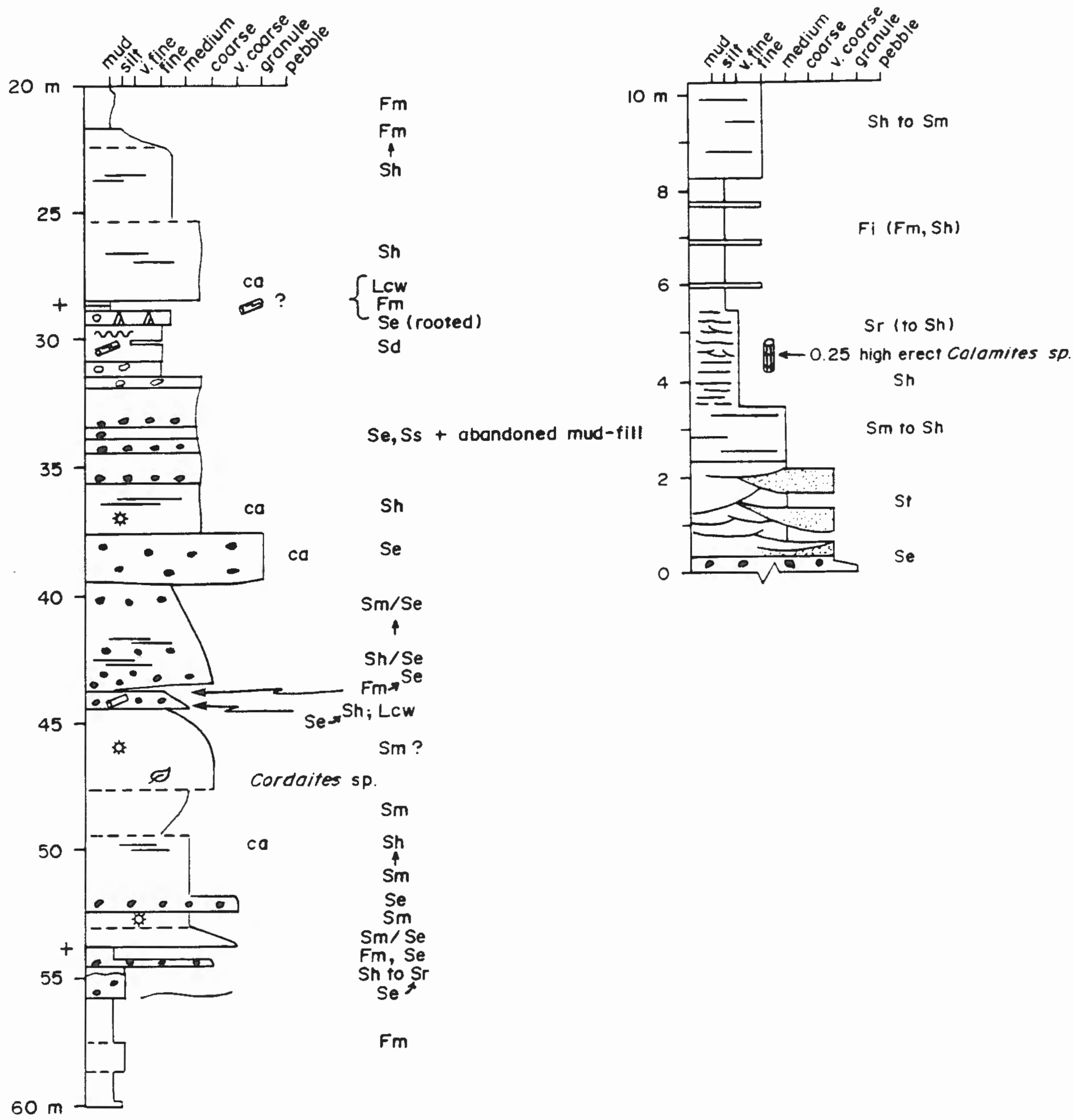


Figure 2.14 a) Vertical profile of 34 m thick multistorey lithic sandstone body, lithofacies assemblage VI, drillhole SH5.

b) Measured section, multistorey lithic to quartzose sandstone body, Harrison Brook.

2.10.3 Interpretation:

The red mudrock/lithic sandstone assemblage consists of fluvial and floodplain deposits and in this respect is similar to the underlying/older variegated and grey mudrock/multistorey sandstone assemblages. This assemblage, however, reflects a history of increased Eh (oxidation levels), a continuation of a trend evident in the underlying variegated lithofacies assemblage. This trend to increased levels of oxidation is attributed to lowering of the groundwater table (cf. Bown and Kraus, 1981), perhaps due to evolution of drainage patterns within the basin or to continued climatic changes (see Chapters III and IV for further discussions).

The paucity of exposure of the multistorey sandstone bodies restricts the extent of interpretation of fluvial style. Although it is somewhat risky to interpret fluvial style on the basis of core profiles, certain fluvial processes and conditions can be surmised from Figure 2.14. The presence of sideritic mudrock (Fm) at various locations in the vertical profile provided by the SH5 drill core suggests a distinctly ephemeral/fluctuating history or flow, similar to the Rodney Sandstone. The abundance of labile constituents may be related to an acceleration of weathering processes enhanced by elevated Eh levels. As previously determined, mudrock intraclasts within the Rodney Sandstone have been shown to be at least partially derived from meso-scale mudrock hollow-fills within the channel belt. In a scenario where fluctuating water levels prevailed, however, bank collapse (Nanson *et al*, 1986) may have been a contributing factor. The saturated muds within these abandoned channel tracts were in places colonized by plants. Permineralized tree axes may represent *in situ* roots or allochthonous trunks and/or roots.

The increase in sandstone body thickness, the continuation of in-channel processes through time at a given reference point, and well-drained overbank sediments, while not conclusive, nonetheless point to a relatively stable, entrenched channel tract. It is worthy of note that recent seismic (Bromley and Calder, in press) and palynological (Dolby, 1991) studies correlate these strata with those exposed near MacCarron's River, interpreted as anastomosed river deposits (Rust *et al.*, 1984).

2.11 SUMMARY

The six lithofacies assemblages defined in this study each exhibit distinctive lithological features arising from depositional processes prevalent within a specific geomorphic region. They further reflect different groundwater levels and position within the basin. The one possible common denominator was climate, but even this factor had arguably changed during deposition of assemblages V and VI (see discussion in Chapter V). The major processes and interpretation of geomorphic setting are summarized in Table 2.6.

LITHOFACIES ASSEMBLAGE	REPRESENTATIVE LITHOLOGY	MAXIMUM KNOWN THICKNESS	MAJOR DEPOSITIONAL PROCESSES/CONDITIONS	PALEOGEOMORPHIC REGION
VI. Reddish mudrock/lithic sandstone	Reddish mudrock; grey multistorey sandstone; calcite nodules	>144m (north limb, Springhill anticline	Streamflow within vertically aggrading fluvial tracts subject to common channel abandonment; overbank deposition under increased Eh levels; immature calcrete formation.	Well drained alluvial plain (anastomosed rivers?)
V. Variegated mudrock/multi-storied sandstone	Mottled mudrock; grey multistorey sandstone; rare coal; calcite & siderite nodules	170m (north limb, Springhill anticline)	Streamflow in well-defined channels; overbank sediment-ation from suspension; rare to common <u>in situ</u> peat accum-ulation; pedogenesis(incipient calcrete formation; early diagenetic redding of mudrocks); elevated Eh.	Alluvial plain similar to above but with fluctuating lower groundwater level.
IV. Mudrock/multi-storied sand-stone (cf."Coal measures")	Grey siderite-bearing mudrock; grey multistorey sandstone bodies; thick (<4.3m) humic coal seams	510m (north limb, Springhill anticline)	<u>In situ</u> accumulation of humic plant matter; fluctuating stage mixed-load streamflow in well-defined channel tracts; overbank sedimentation from suspension; precipitation of carbonate minerals in saturated soil or quiet water; widespread reducing conditions	Poorly drained alluvial plain; forested peat mires; medial reaches of major fluvial channel belt; well protected, perenially saturated flood basin.
IIIB. Bivalve-bearing association of assemblage III	Bivalve-bearing shale & limestone; humic to sapropelic coal (often impure)	>10m north limb, Springhill anticline)	Siliciclastic and bioclastic deposition in partially oxic lake waters; repeated discrete overbank/interdistributary flooding; deposition of humic to sapropelic peat.	Vegetated freshwater lake margin with low-lying soligenous peat mires and fluvial channels.
III. Thinly interbedded	Thinly interstratified sandstone and mudrock; rare, thin quasi-sapropelic coal	173m (north limb, Springhill anticline)	Predominantly suspension and density-flow deposits; deposition of allochtonous organic matter; local subaerial exposure.	shallow lake/bay.
II. Poorly sorted, gradationally interbedded	Poorly sorted, crudely bedded granule conglomerate and mudrock	>44m (south limb, Springhill anticline)	Ephemeral, channel and sheetflow; possible dilute mudflow. Early stages of pedogenesis.	Down-fan regions distal to deposits of conglomerate assoc'n at or near fan toes.
I. Conglomerate	Stratified, frame-work/sandy matrix orthoconglomerate	>250 (Polly Brook)	Fluctuating stage streamflow; gravelly bars. (Debris/ mudflow in some regions outside study area).	Mid-distal alluvial fans, bajada/piedmont plain.

Table 2.6 Summary of lithological, depositional and paleoenvironmental aspects of lithofacies assemblages, Springhill Coalfield, south-central Cumberland Basin.

CHAPTER III: BASIN ANALYSIS:
ALLOGENIC CONTROLS ON ANCIENT PEAT FORMATION

3.0 INTRODUCTION

The application to coal basin analysis of the theory that the location of peat-forming ecosystems is "determined by the underlying topographic and hydrologic patterns of that landscape" (Tallis, 1983), is the focus of this chapter. Broader conditions for the existence of a coal-forming peat swamp were specified by Teichmüller (1982) as: i) evolutionary development of the flora; ii) climate; and iii) geography and structure of the region, including: a) slow continuous groundwater rise paralleling peat accretion provided by subsidence; b) protection against major marine or terrestrial incursions; and c) a low base level and restricted siliclastic sediment supply. In the consideration of the place of peat mire development during deposition of the Cumberland Group in the Cumberland Basin, the Westphalian B time frame precludes any major evolutionary change in the flora. The two remaining factors, climate and geographic/tectonic setting, are reflected in hydrology and topography as suggested by Tallis (1983). This chapter addresses the hypothesis that topography and hydrology, which are largely allogenic factors in mire development, determined locations conducive to mire formation in the Cumberland Basin.

Wetland ecosystems may develop in response to allogenic and/or autogenic change. Allogenic changes are those produced by influences external to the mire (Gore, 1983) including tectonic, geomorphic and climatic factors. Autogenic change results from processes internal to the mire (Gore, *ibid.*) such as floral succession and internal drainage development (Tallis, 1983). These principal categories of influences on wetland development should be applicable to ancient peat mire systems, the precursors of coal seams.

This chapter deals with the allogenic influences of paleogeography on mire formation in the Cumberland Basin. In the preceding chapter, the sedimentology of lithofacies assemblages from the Springhill coalfield was appraised. From this, the environments and geomorphic settings of deposition were interpreted. In this chapter the assemblages are placed in a stratigraphic context from which it is possible to infer temporal and spatial aspects of the paleogeography. By identifying tectonic and climatic controls on the evolution of the basin-fill, it is then possible to interpret allogenic constraints on the development of the peat mires. Autogenic and allogenic change in the development of one of these ancient mires is the focus of Chapter IV.

3.1 STRATIGRAPHY OF THE SPRINGHILL COALFIELD

3.1.1 Lithostratigraphy: Architecture of Lithofacies Assemblages

Due to the sporadic nature of outcrop in the region of the Springhill coalfield, documentation of the stratigraphic arrangement of lithofacies assemblages draws heavily upon drillhole correlation and especially the correlation of coal seams. These sedimentary bodies lend themselves to such use due to their areal extent, their distinct geophysical signatures recorded through borehole logging, and because some seams on the northern limb of the anticline were extensively mined in the subsurface.

The correlation of seams throughout the coalfield was only recently achieved (Calder, 1980) and verified by subsequent diamond-drilling (e.g. Calder, 1980, 1981a, 1981c, 1982). Historically, the main hindrances to correlation of seams were the structural complexity of the coalfield (chiefly the abundant normal faults subparallel to the anticlinal axis), abrupt lateral lithologic change (largely coincidental with the highly deformed axis), and the lack of detailed sedimentologic analysis of the strata. These factors were compounded by the paucity of outcrop in the region.

The inter-borehole correlation of coal seams and lithofacies assemblages along a North-South transect of the coalfield is portrayed in Figure 3.1 (in rear pocket). This chart is the basis of the structurally reconstructed stratigraphic section of the Springhill coalfield depicted in Figure 3.2, using the No. 3 seam as datum. The two sections illustrate the spatial relationships of lithofacies assemblages and position of major coal seams. (Note that there is a 2:1 exaggeration of vertical scale in Figure 3.2 so that individual coal seams may be distinguished).

Although the lithofacies assemblages of this study are defined on the basis of sedimentology, it is obvious from Figures 3.1 and 3.2 that they also occur as discrete bodies within the basin. As such, the assemblages fulfil the broad criterion set forth by the North American Commission on Stratigraphic Nomenclature (1983, Article 25a) for the formal designation of members, namely that "a member is established when it is advantageous to recognize a particular part of a heterogeneous formation". The requirement of mapability is not mandatory for the designation of a member (Article 24d) although the associations-cum-members of this study are indeed mappable within the 100 km^2 area of the geological map of the Springhill coalfield (Calder, 1990). Many of these lithofacies assemblages have since been adopted as formal lithostratigraphic units (Ryan *et al.*, 1990; in press).

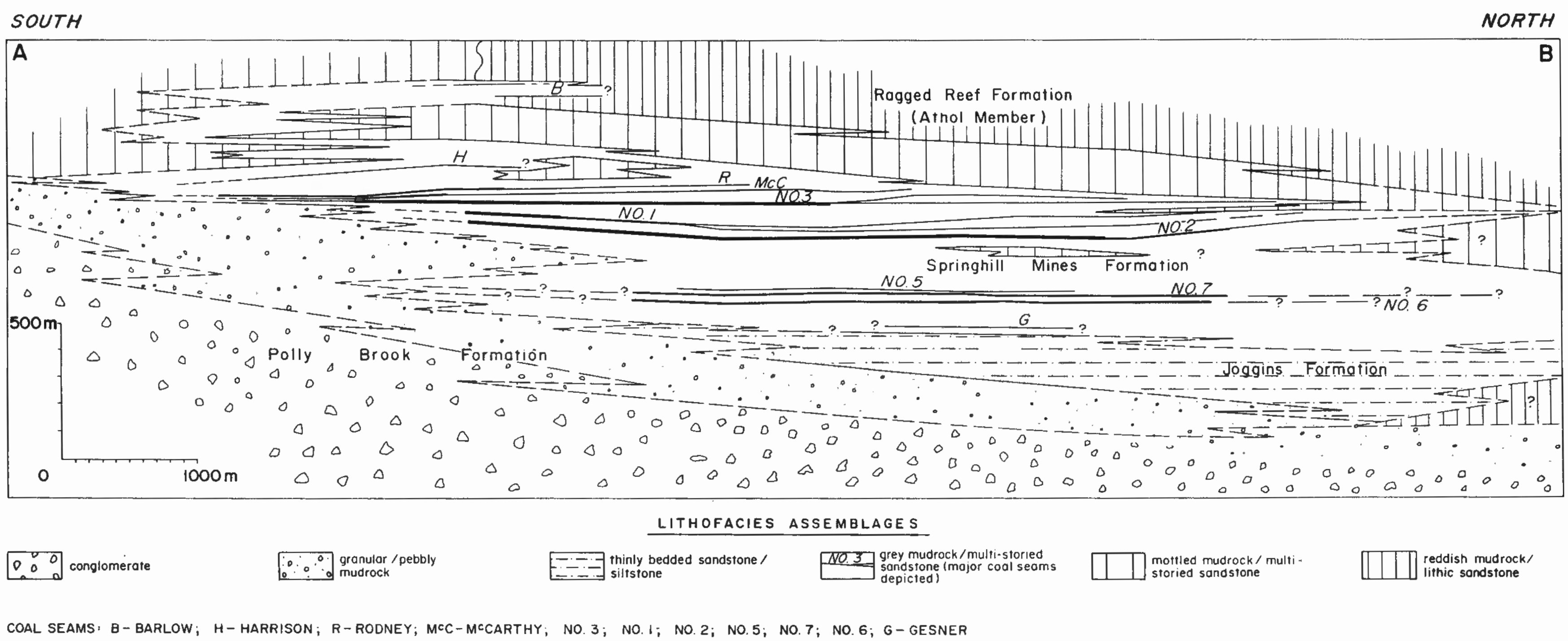


Figure 3.2 North-south stratigraphic section, Springhill coalfield

The precise nature of the stratigraphic boundaries between various member assemblages is not well understood in all cases. The relationship between the coal measures of assemblage IV of the Springhill Mines Formation (Figs. 3.1 and 3.2) and those with which it is in direct contact, especially II and V, are comparatively well documented. (An exception is the stratigraphic relationship between the Springhill Mines Formation (assemblage IV) and the Joggins Formation (III), the boundary having been defined only in drillhole SH74).

In contrast, the spatial relationship between assemblages I and II, within the Polly Brook Formation, is largely inferred. Field mapping in the Rodney-Leamington area in the south of the coalfield suggests a gradational contact, perhaps with some intercalation as indicated along Black River north of the Black River Road. On the northern limb of the Athol syncline at Styles Brook, the Leamington Member (assemblage II) is virtually absent. The distribution of assemblage II is not well understood north of the Springhill Anticline, although it appears to become much reduced in thickness, eventually pinching out.

The Leamington Member (assemblage II) underlies and interfingers with the Joggins Formation (assemblage III), as evident in drillhole SH74, however, little else is known of their spatial arrangement. The contact between the Leamington Member (assemblage II) and the "coal measures" of the Springhill Mines Formation (assemblage IV) has been intersected in numerous drillholes along and south of the axis of the Springhill Anticline. The complex interdigitation of these two formations and assemblages is also evident on Black River north of Black River Road, and along Boss Brook between Rodney and Polly Brook. The thinly intercalated nature of these units (Figures 3.1, 3.2) requires that their contact be averaged when drawn in plan (map) view.

The coal seams within assemblage IV of the Springhill Mines Formation provide marker beds which permit a clearer understanding of the stratigraphic relationship between the coal measures and assemblage II, the Leamington Member of the Polly Brook Formation. Even so, this relationship is poorly understood for strata below the No. 2 seam, as reflected by the dashed lines in Figures 3.1 and 3.2. This is due to the lack of available drill core intersections of this interval and to the fact that these strata subcrop in a particularly deformed region of the coalfield with little outcrop.

Within the Springhill Mines Formation the grey coal-bearing assemblage IV is overlain by unassigned strata of the reddish mudrock-bearing assemblage VI. The contact between the two is essentially transitional from a predominantly grey mudrock with major coals to a predominantly red mudrock virtually devoid of coal. This lithologic transition is represented by lithofacies assemblage

V, the MacCarrons Creek Member, which has been assigned to the Springhill Mines Formation by Ryan *et al.* (1990; in press). This assemblage occurs as isolated tongues and lenses particularly above the Rodney seam horizon. In the north (Figures 3.1 and 3.2) and to a lesser degree southeast of the coalfield, these mottled strata occur lower in the section, lateral to the uppermost major seams (No. 3, McCarthy, Rodney). Above the Rodney seam horizon, tongues and lenses of the MacCarrons Creek Member (assemblage V) become widespread. In the region of the anticlinal axis (south-central), the digitation of grey strata (assemblage IV) and mottled/variegated reddish strata within the Springhill Mines formation is most pronounced, with grey, coal-bearing strata persisting as high as the Barlow seam horizon (Figure 3.2), whereas laterally equivalent strata to the north and east are virtually entirely of the mottled assemblage V.

3.1.2 Biostratigraphy and Age of the Springhill Coalfield

The basin-fill sequence, comprising the entire Springhill Mines Formation (assemblages IV, V), the underlying Joggins Formation (assemblage III) and overlying beds (assemblage VI), assigned to the Athol Member of the Ragged Reef Formation was systematically sampled by the writer and submitted to Dr. G. Dolby, Calgary, for palynological analysis; these and subsequent sample locations are indicated on Figure 3.1 (see also Figure 3.3 in rear pocket). A vertical sampling interval of 30 m was employed where possible. Coal and siliciclastic lithofacies were sampled separately. These lithofacies together provide a broad spectrum of palynomorphs. Impure coal (coaly shale) typically yielded a diverse palynoflora that was particularly useful for biostratigraphic determinations. The coal lithofacies are important in that the contained palynomorphs are considered to be essentially *in situ* with minimal reworking (Traverse, 1988). In order to minimize environmental effects on the spore assemblage within the siliclastic suite, care was taken to sample similar lithofacies throughout: grey, fine mudrock, commonly with siderite and compression flora (lithofacies Fm, predominantly). In all cases the lithofacies was recorded. Palynological analysis is constrained to a large degree by the occurrence of grey argillaceous strata (lithofacies Fm, Fc and Fl) and coal which yield abundant miospore assemblages. Consequently, the age of strata of the conglomeratic assemblage I and of the reddened mudrock assemblages V and VI can only be determined approximately, relative to the intervening strata of assemblages III and IV.

The entire 1050 m-thick succession that was sampled was initially assigned a late Westphalian B age (Dolby, 1984). Coal-bearing strata at Springhill have been previously assigned a Westphalian B age on the basis of macroflora (Bell, 1944) and palynoflora (Hacquebard and Donaldson, 1964).

Exhaustive sampling and analysis of the Pennsylvanian strata within the Cumberland Basin and in particular along the entire coastal section has permitted the resolution of seven spore zones within the Westphalian A and B (Dolby, 1987; 1991). These subsequent data, including detailed study of Springhill coals for the purpose of paleoecological reconstruction (Dolby, 1988a; Calder, in press) and resampling of problematic sections at Springhill, has led to a revised age of late Westphalian A to Westphalian B for the coal-bearing section of the Springhill coalfield (Dolby, 1991). The precise Westphalian A-B boundary is poorly defined even in the European type areas, but is thought to occur between the No. 7 and No. 2 seams (Dolby, 1991).

Zones 3 to 4 of Dolby are represented in the Springhill coal-bearing section. The Nos. 7 and 6 and Gesner seams, including the lowermost part of the Springhill Mines Formation (lithofacies assemblage IV) and the Joggins Formation (lithofacies assemblage III), are assigned to Zone III, of late Westphalian A age (Dolby, 1991). *Secarisporites remotus*, found within the No. 7 seam, dies out at the top of the Morrowan in the Illinois Basin (Peppers, 1985), and in the mid-Westphalian A or lower part of the RA zone of Clayton *et al.* (1977) in Europe (Dolby, 1991). Zone 4A (latest Westphalian A) is poorly defined within the sample suite, but the No. 2 seam (Springhill Mines Formation; lithofacies assemblage IV) yielded specimens of *Punctatosporites spp.*, *Lophotriletes "spinosaetosus"*, and *L. ibrahimii*, diagnostic of Zone 4B (earliest Westphalian B). An occurrence of *Florinites junior* within the No. 3 seam has been used to mark the base of Zone 5 (Westphalian B) (Dolby, 1991). No data exist for the basal conglomerates of the Polly Brook Formation (assemblages I and II) but samples from the underlying Boss Point Formation at Springhill have yielded Zone 2 (early Westphalian A) spore assemblages (Dolby, 1988b; 1991). Deposition of these conglomerates is presumed to have taken place in the main during the Westphalian A, but with minor deposition extending into the Westphalian B.

3.2 TEMPORAL AND SPATIAL ASPECTS OF THE STRATIGRAPHIC RECORD AT SPRINGHILL

3.2.1 Architectural Development of Coal Seams

As described in Chapter II, the major coal seams of the Springhill coalfield occur within lithofacies assemblage IV (Springhill Mines Formation), which is additionally characterized by grey mudrocks and multistorey sandstone bodies. This relationship is apparent in Figures 3.1 to 3.2. Nearly 90 percent of aggregate coal thickness (48 m) and >80 percent of seams greater than 5 cm in thickness occur within lithofacies assemblage IV.

Within the Springhill Mines Formation, 15 coal seams have been identified that locally attain a thickness greater than 1 m. Seven of these, constituting the major seams of recent economic interest, are depicted in (north-south) cross-section in Figure 3.4. These cross-sections are, in effect, transverse sections of the seams that include, in some cases, their depositional boundaries. The sections of the upper three seams (Rodney, McCarthy and No. 3, in descending order) represent a nearly complete transect from the southern to northern seam margins. The sections of the remaining four, (No. 1, No. 2, No. 7 and No. 6) are to varying degrees incomplete. The Nos. 7 and 6 seams in particular extend for an undetermined distance beyond that depicted in Figure 3.4, although all show some degree of splitting at the ends of the sections. The transverse distance spanned by these seams varies from 4.5 km (Rodney seam) to 6.5 km (No. 3 seam) or greater (at least 7.5 km: No. 2 seam).

The longitudinal extent of the seams remains largely unknown. The eastern extension of the westerly-dipping seams is not known due to erosion of the coal-bearing strata in the vicinity of the Black River diapir. The major seams in general do not exhibit any signs of deterioration in their easternmost locations, near their outcrop/subcrop. The western limits of the coals are similarly unknown, due to the great depths which the coal seams reach. Even at a vertical depth of 1323 m, the No. 2 seam was reported to maintain a high degree of development with relatively low ash yield (Copeland, 1959), although isopach maps (Copeland, *ibid.*) arguably signal the start of closure of the extent of the No. 2 seam at around this level. The known areal extent of a seam may be greater than 40 km^2.

Coal seams also exhibit a pattern of areal distribution that varies systematically with stratigraphic position (Figures 3.2 and 3.5). The major seams, older than and including the No. 2 seam, which occur within the lower part of assemblage IV do not appear to show a clear pattern of change in areal development. Above the horizon of the No. 2 seam, however, younger seams encroach progressively further southward. The younger coal seams (No. 2 - Rodney interval, perhaps including the Harrison seam) therefore record a progressive southward onlap in conjunction with the onlap of the host grey mudrock/multistorey sandstone lithofacies assemblage (IV) of the Springhill Mines Formation over assemblage II of the Polly Brook Formation. This seam interval (No. 2 to Rodney/?Harrison) also shows the maximum thickness of coal and multistorey sandstones. The significance of this relationship will be discussed later.

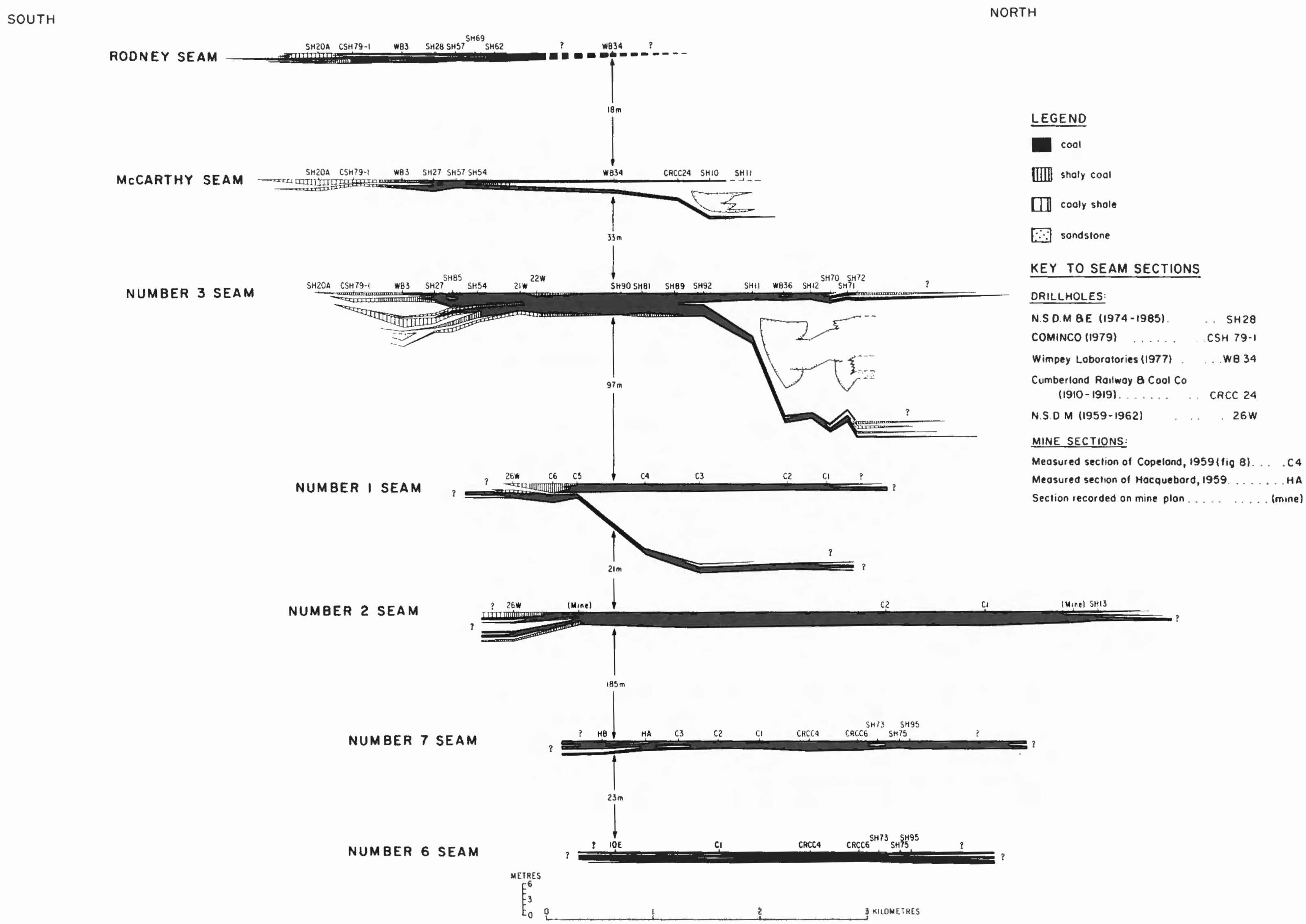


Figure 3.4 Cross-sections (N-S) of major coal seams of the Springhill coalfield.

3.2.2 Geometric and Lithologic Zonation of Seams

The major coal seams (Fig. 3.4) exhibit similar geometry and modes of lateral termination (see Table 4.2, Chapter IV). All possess *inner mire* regions devoid of major siliciclastic partings and within which dispersed inorganic matter and sulphur contents are lowest. Not surprisingly, these inner zones were the preferential sites of mining in the past, barring structural complexities. Both the northern (riverine) and southern (piedmont) seam margins commonly exhibit extensive splitting and interdigitation with siliciclastic sediments. There are, however, lithologic and geometric dissimilarities between the two margins, some of which are readily apparent even at the scale of the seam cross-sections in Figure 3.4.

The *piedmont* margins of the coals exhibit abrupt southward gradation from low to high ash coal, with a concomitant development of intervening siliciclastic partings. The point at which successive partings within a given seam pinch out to the north or lose their discrete identity tends to be vertically coincidental within that seam. This stacking of parting "hinge-lines" results in an abrupt deterioration in coal quality and a thickening of the seam to its maximum over a short distance. So abrupt is this "internal margin" within some seams that it led to the interpretation of a "wash-out" of the No. 2 seam (Copeland, 1959), an idea that prevailed until recently (Calder, 1981c). The rapid southerly decrease in organic content of the seams results in a laterally abrupt transition from coal (< 25% ash by weight) through impure coal (> 25% but < 70% ash) to carbonaceous mudrock. This phenomenon was recognized by Hacquebard *et al.* (1967) who referred to it as "lithification". Hacquebard and Donaldson (1969) later used this attribute, among others, to distinguish limnic from paralic conditions. The author concurs with their observation but suggests usage of the term intermontane in lieu of limnic so as to avoid confusion between the terms limnic and lacustrine.

Siliciclastic partings at the piedmont margin include abundant carbonaceous mudrock (Fc) similar to partings of the riverine zone, however laminated or crudely stratified siltstone and fine-to coarse-grained sandstone (Sd, Sg, Fm, Fl, and Shl) are common. The strata enclosed by the coal commonly coarsen upward and the sediments are poorly sorted. Crudely stratified carbonaceous matter is common within the mudrocks (Fcl). Bark vitrain bands and lenses are common within both mudrocks and sandstones, as are roots.

The *riverine* margins of the coals exhibit more geometric variation between seams than is the case with the piedmont zone. The seams generally experience a more gradational northward increase in ash content, so that each leaf of the seam maintains an inner-zone macroscopic character for a much greater distance (up to 2 km) than do leaves in the same seam at the piedmont margin (Figure 3.4).

A characteristic unique to the riverine margins of the seams is the profound splitting of some seams (e.g. McCarthy, No. 3, No. 1) about multistorey sandstone bodies (Figure 3.4). Where a seam bifurcates around such bodies, the split exhibits a dramatic northward increase in thickness. Such splits are up to 22 m thick (e.g. No. 3 seam, Figure 3.4). Thinner mudrock partings commonly extend at the same stratigraphic level as the sandstone bodies for a kilometre or more into the inner zone of the seam. In certain seams, such partings extend sufficiently far as to realize a near or complete dichotomy of the seam (e.g. McCarthy and No. 1 seams, Figure 3.4). Less dramatic intercalations (dm- to m-scale) of mudrock at the riverine margins of the seams, without entombed multistorey sandstone bodies, are lithologically similar to those which adjoin the sandstones. They comprise dark grey carbonaceous mudstone (Fc), in places bearing laminar siderite.

3.2.3 Areal Development of Coal Seams

The temporal and spatial development of the major coal seams, including the progressive southerly onlap of seams younger than the No. 2, is further depicted by sequential seam development maps (Figure 3.5). These maps depict the major areal zones of peat (coal) accumulation within the seams, approximately equivalent to the inner zone described above. The maximum zone of peat (coal) accumulation was delineated on the basis of the equation

$$Tc \leq 0.8\ Tc_{max}$$

where Tc_{max} equals the maximum aggregate thickness of coal (less than 25 percent ash, as per ICCP, 1963) and Tc is the thickness of coal at a given location. Tc_{max} invariably occurs in the central region of the seam. This method of defining the maximum zone of peat accumulation is useful in that the long-standing term such as "shaly coal" employed by Nova Scotian coal geologists similarly refers to impure coal having an approximate ash yield of > 25 percent. Mine sections can therefore be utilized by subtracting the thickness of impure coal and inorganic partings. To this end, the isopach contour maps and measured sections of Copeland (1959; his Figure 6, sheets 1-6, and his Figure 8) were particularly useful in defining the aspect of deep-lying seams no longer accessible through active mine workings. Weighted ash yield would doubtless provide a more accurate zonation, but sufficient data are not available for all seams. The limit of eighty percent of maximum coal thickness was chosen

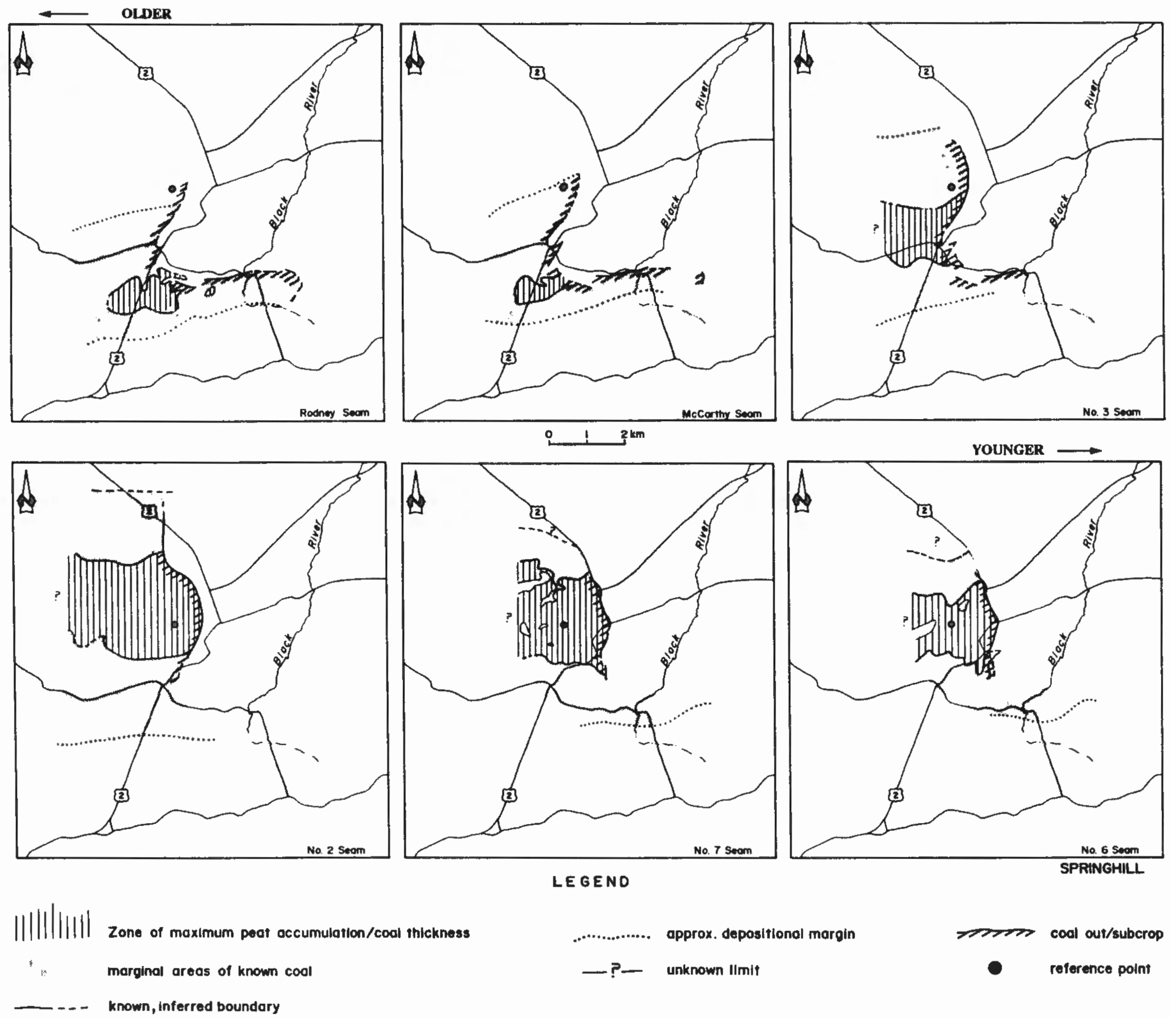


Figure 3.5 Seam development maps, Springhill coalfield, showing inner zone of maximum coal development, and inferred extent of seams. Reference point (solid black dot) for all seams taken from developmental center of the No. 6 seam.

as it was found to coincide approximately with a weighted ash yield of 25 percent for the No. 3 and Rodney seams. Areas of local thinning within the zone of maximum accumulation are indicated accordingly (e.g. Nos. 6 and 7 seams, Figure 3.5). Major constraints on the seam development maps are: 1) absence of thickness data at depth (to the west) within the older, deep-lying seams beyond the limits of underground mining and drilling; and 2) absence of data to the east or northeast of seam outcrops imposed by erosion.

Scrutiny of the spatial configuration of these similarly defined zones reveals that the Nos. 6 and 7 seams exhibit a generally similar pattern of development, with several elongate areas of internal local thinning. The zone of maximum accumulation shows a generally east-west orientation, albeit within the constraints described above. The configuration of the No. 2 seam is similar, but shows a larger expanse and apparently little or no internal thinning.

Major departures in the configuration and location of the zone of maximum accumulation occur in the younger No. 3, McCarthy and Rodney seams. The zone is located progressively further south within each of these seams. The southern limit of the zone for the Rodney seam occurs more than 3 km further south than the limit for the No. 2 seam, which lies approximately 180 m stratigraphically below the Rodney. Consistently thick coal within the zone of maximum accumulation likewise becomes much reduced within successively younger seams above the No. 2. The minor (north-south) axis of this elliptical zone decreases from 2.5 km in the No. 2 seam to 1 km for the Rodney seam.

3.2.4 Spatial Reconstruction of Lithologic Entities

Further paleogeographic reconstruction can be achieved by considering the position of coeval deposystems (lithofacies assemblages of Chapter II) with respect to the peat mires (coal seams). The No. 3 seam was chosen for this reconstruction because of the relative abundance of data.

Figure 3.6 depicts the spatial arrangement of lithosomes at the approximate mid-history of the No. 3 seam, when major incursions occurred in the piedmont and riverine zones. The incursions may not have been precisely coeval. Lithofacies assemblages V and VI occupy the most distal/basinward (northerly) position. Coal-bearing assemblage IV occupies a medial position, and lithofacies assemblages I and II border the coal-bearing strata on their southern, proximal margin. Although it is clear that basal conglomerates of assemblage I occupy a proximal position adjacent to the basin margin, the spatial relationship between assemblages I and II is an approximation. Closure

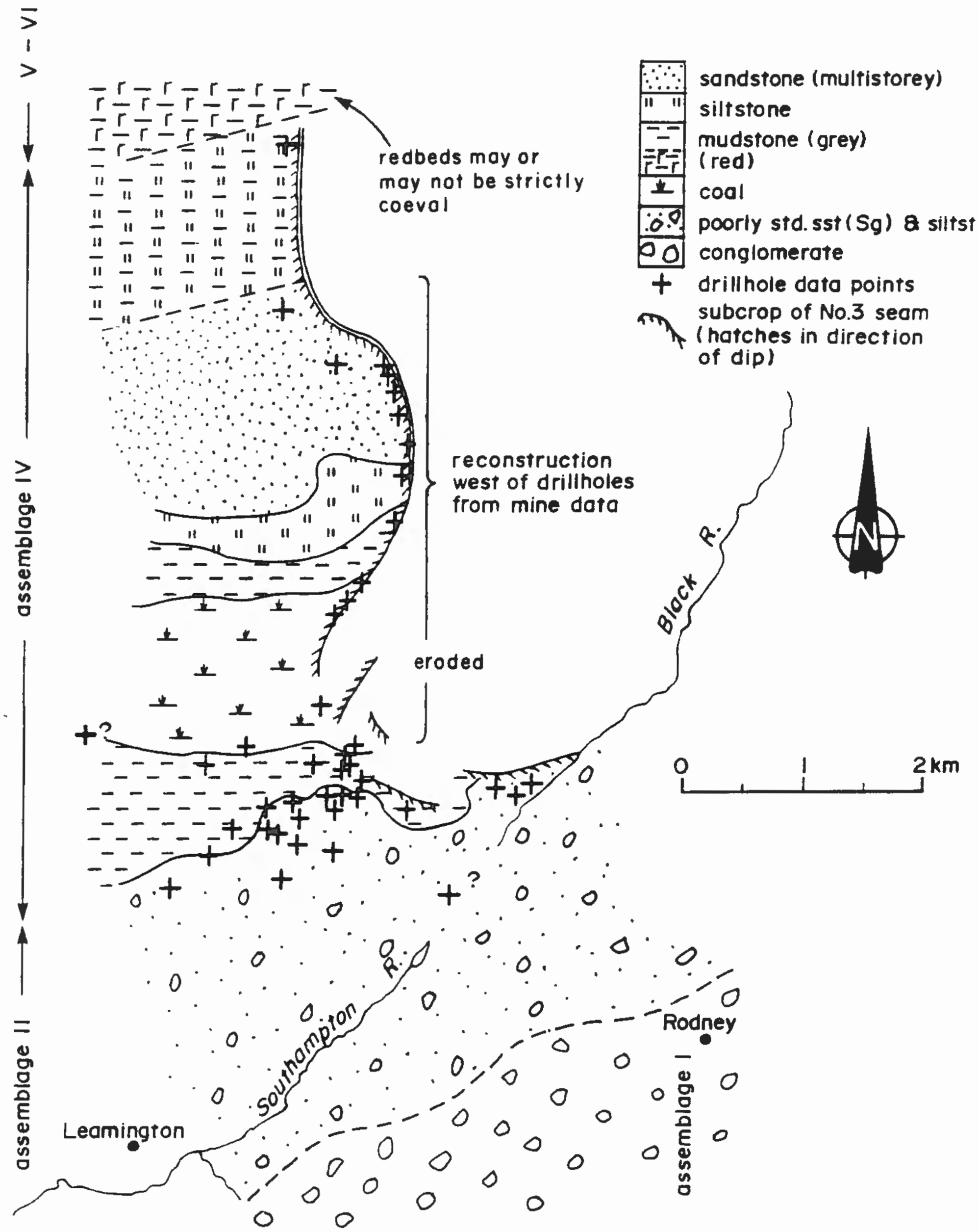


Figure 3.6 Areal map of No. 3 seam and surrounding lithosomes, representing period during which major (approximately coeval?) seam splits developed in the north and south, resulting in near-dichotomy of seam.

of the systems to the west is not shown due to the absence of data at depth.

A kilometre-wide multistorey sandstone body occupied a position at the northern margin of the coal seam. Basinward of this sandstone lithosome reddish mudrocks occur. The coal lithosome for the time interval depicted coincides approximately with the zone of maximum thickness/accumulation (Figure 3.5). Marginal areas of the seam, both north and south, were at this time overlain by the bordering lithosomes (see Figure 3.6). To the south of the No. 3 seam, a sinuous margin existed with a grey mudrock and a more proximal ill-sorted sandstone lithosome, the latter typified by a coarse, graded sandstone lithofacies (Sg) of assemblage II. Further south, the basin margin was occupied by extra-formational conglomerates of assemblage I.

3.3 PALEOFLOW ANALYSIS

Paleoflow indicators were measured during mapping of outcrop and the database is consequently constrained by the occurrence and nature of the outcropping strata. The types of paleocurrent indicators measured are ripple marks on bedding planes, ab plane dips (imbrication) of granules/pebbles, plant stem long-axis orientation, large-scale crossbedding (both trough and planar), flute and groove casts, and meso- to macro-scale hollow-fills. Analysis of lithosome orientation from mine and closely spaced drillhole data provides important data of a larger scale. In order to interpret these paleocurrent data in the light of paleodrainage patterns and hydrodynamic processes, it is necessary to consider 1) the sedimentological hierarchy of the indicators, and 2) the relative quality of data measured.

3.3.1 Hierarchy of Sedimentary Structures

Allen (1966) considered that flow and the structures subsequently produced represent five orders of magnitude with 0, the largest scale of paleoflow indicator, representing an individual river and 4, the smallest, representing small-scale superimposed structures such as ripple-drift cross-stratification. Miall (1974) quantified the directional variance of the structures within each rank and expanded Allen's ranking system from five to six. Miall added as the highest rank the entire river system, which was alluded to by Allen but not included in his enumeration. A comparison of these two ranking schemes is presented in Table 3.1, along with the types of sedimentary structure/paleocurrent indicator measured in this study.

RANK		TYPE OF SEDIMENTARY STRUCTURE	SPECIFIC PALEOCURRENT INDICATOR (This Study)
ALLEN 1966	MIALL 1974		
0	2	Meander belts of individual rivers	Isopachs of channel sandstone bodies from drillholes or mine workings (I)
1	3	Major channel reaches within individual meander belts	Macro-scale mudrock hollow-fill (D)
2	4	Minor subsidiary channels; bars	Meso-scale hollow-fills; large scale lateral accretion sets (D)
3	5	Structures within bars; small-scale erosional features	Trough and planar cross-bedding; flutes, grooves (D)
4	6	Structures superimposed on larger-scale structures	Ripple cross-lamination; pebble orientation; fossil orientation; parting lineation (D)

Method of Measurement: D = Direct; I = Indirect

Table 3.1 Hierarchy of sedimentary structures in fluvial deposits after Allen (1966) and Miall (1974) with examples of paleocurrent indicators measured in this study, south-central Cumberland Basin.

In his study, Miall (1974, his Figure 2) showed that the range of directional variance shown by sedimentary structures increases from highest rank (his 1) to lowest rank (his 6). As one would expect, the largest portions of the river system best reflect the orientation of that system. Sinuosity of the river system, only briefly touched upon by Miall, is expected to have a pronounced effect on the directional variance of major channel-reach segments (Miall's rank 3). In high sinuosity systems, one or two measurements of this type may be highly misleading. The greatest increase in variance is between Miall's ranks 4 and 5. Although structures of ranks 5 and 6 show a high range of directional variance, these are the types of sedimentary structure most often available for paleocurrent measurement (Miall, 1974). Accordingly, a greater quantity of data for troughs or ripples is required to obtain an accurate mean vector than is required for higher rank structures.

3.3.2 The Paleoflow Record at Springhill

In all, 183 paleoflow measurements were recorded during the course of this study. Summary rose diagrams of these data are given in Figure 3.7. The majority (56 percent) of data were recorded from lithofacies assemblage IV, representing five ranks within the hierarchy of indicators (Table 3.2). Note that when reference is made in the text to flow direction, this refers to downstream direction (i.e. northeasterly flow means flow to, not from, the northeast).

3.3.2.1 Lithofacies assemblage I: basal conglomerates

Paleocurrent indicators within the basal, extra-formational conglomerates (Figure 3.7) are uncommon, with only 6 large-scale troughs measured. Trough axial trends show northwest/southeast orientation and the few sufficiently well exposed in three dimensions (n=3) indicate a northwesterly flow direction. Such a paleoflow direction is also suggested by areal fining/thinning trends (cf. Blissenbach, 1952; Bluck, 1964; Boothroyd, 1972; Kesel and Lowe, 1987) within the assemblage in the vicinity of the Springhill coalfield. The trend of the conglomerate lithosome is considered similar to rank 1 of Miall.

3.3.2.2 Lithofacies assemblage II

The poorly sorted strata of assemblage II yielded a weakly bimodal flow pattern with a northwest trend predominating over a lesser north-northwest trend (Figure 3.7). Unidirectional data indicate a northwest early paleoflow. All measurements from trough cross-stratified conglomerate (Gt) fall within the NW quadrant, whereas only 17 percent of measurements from

LITHOFACIES ASSEMBLAGE		RANK	SEDIMENTARY STRUCTURE	NO.
I	Conglomerate	5	large-scale troughs	6
II	Poorly Sorted	5	Meso-scale hollow-fill	1
IV	Coal-bearing	2	Channel-belt isopachs	3
		3	Macro-scale hollow-fill	14
		4	Meso-scale hollow-fills	13
		5	Large-scale troughs	3
		6	Ripple cross-stratification, parting lineation; plant stems; granule trains	75
VI	Red Mudrock	4	Meso-scale hollow-fill	2
		5	Large-scale troughs	8
		6	Ripple cross-stratification; parting lineation	41
			TOTAL:	183

Table 3.2 Breakdown of paleocurrent database by lithofacies assemblage and rank of sedimentary structure

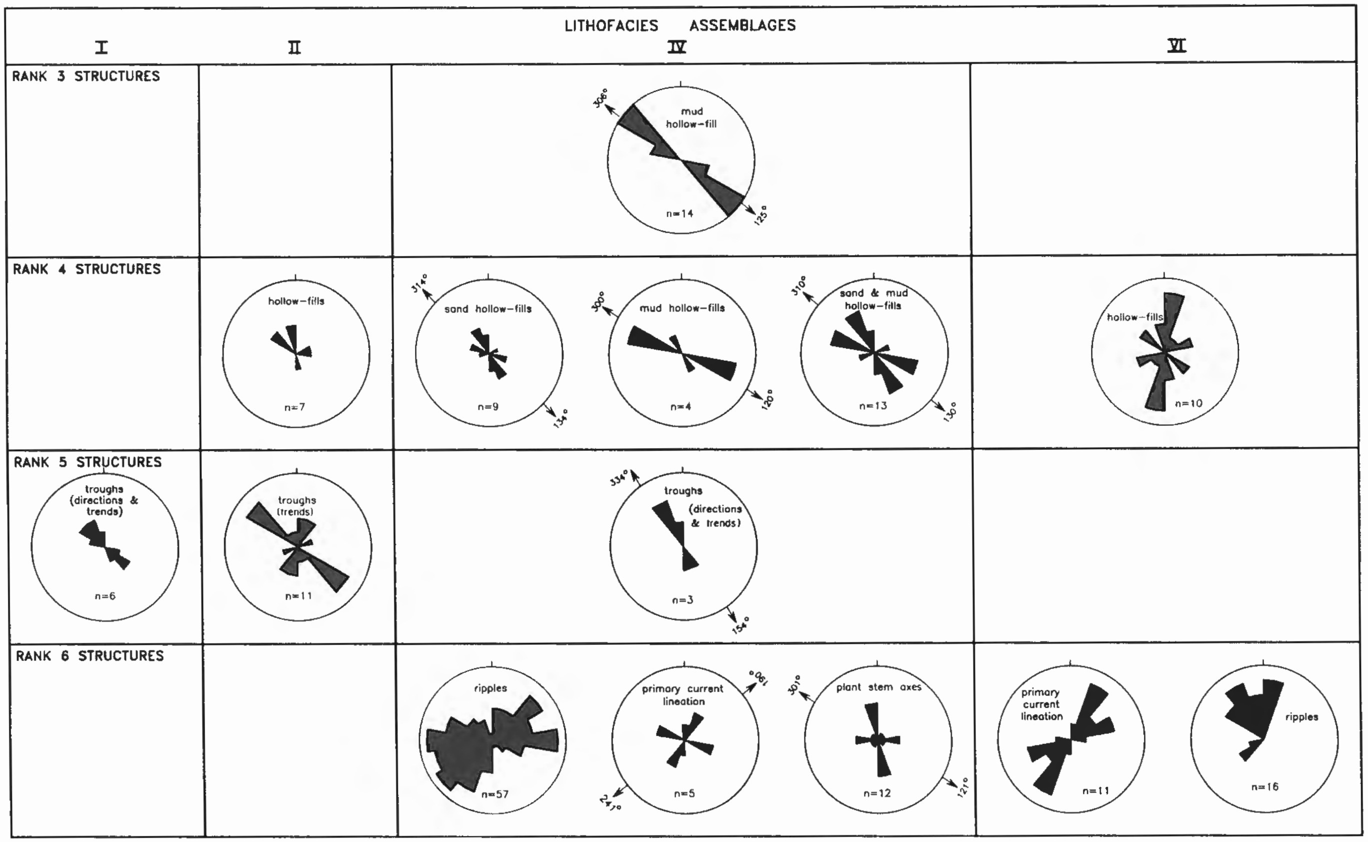


Figure 3.7 **Paleoflow indicators measured for all lithofacies assemblages, Springhill coalfield.**

trough-cross-stratified sandstone (St) lie within the NW quadrant, the remainder lying in the two eastern quadrants.

3.3.2.3 Lithofacies assemblage IV: coal measures

The bulk of the measurements were recorded within the Novaco open-pit mine during its excavation. The orientation of three multistorey sandstone bodies, two enveloped by northerly splits in the Nos. 3 and 1 seams, and one lying above the Rodney seam, provide important data of high rank (2 of Miall, 1978). The three bodies trend northeast to east (Figure 3.8).

The Rodney Sandstone (see Chapter II) provides further paleoflow data (Figure 2.10) spanning the spectrum of rank from Miall's 2 to 6. Mudrock-rich, macro-scale hollow-fills and meso-scale hollow-fills, of ranks 3 and 4, respectively, yield a strong northwest/southeast vector azimuth. These, however, are nearly perpendicular to the overall channel belt orientation, as indicated by the trend of the multistorey sandstone body (Figure 3.8).

Lower rank (5 and 6) structures from within the Rodney Sandstone show a varied orientation. The few trough axes recorded trend northwest/southeast. Coalified stems of *Cordaites sp.* and other plants show a north-south axial orientation. Parting (primary current) lineation is strongly bimodal with northeast/southwest and northwest/southeast modes. Ripple cross-stratification, primarily of the climbing ripple type, shows a southwest unidirectional vector mean within the Rodney Sandstone (Figure 2.10).

3.3.2.4 Lithofacies assemblage VI: reddish mudrock

Measurements from lithofacies association VI (Figure 3.7) were obtained primarily from the vicinity of Athol, west-northwest of the coalfield (Figure 1.3). Rank 4 and 5 data show a north-northeasterly trend. Primary current lineation (rank 6) exhibits a similar northeast/southwest orientation. Rank 6 ripple cross-stratification however indicates predominantly north-northwesterly flow.

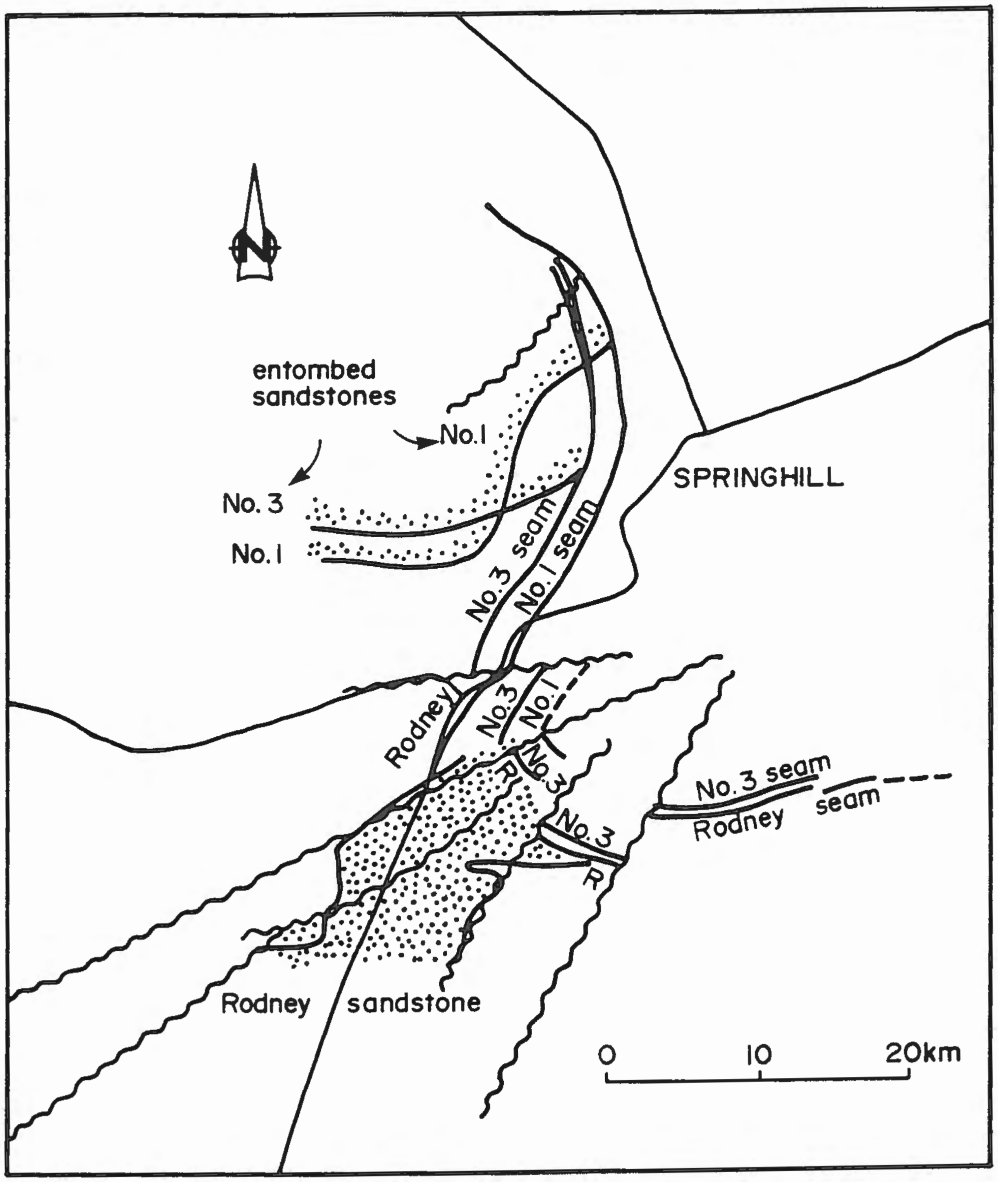


Figure 3.8 Trends of multistorey sandstone bodies entombed within the Nos. 1 & 3 seams (southern margins) and overlying the Rodney seam. Note similar alignment of major fault trends.

3.4 PALEOGEOGRAPHY OF THE SPRINGHILL COALFIELD

The paleogeographic setting of the coalfield can be interpreted by combining:

1) identification of deposystems from the sedimentology of lithofacies assemblages;
2) spatial reconstruction of deposystems from stratigraphic analysis;
3) temporal considerations from coal seam architecture; and
4) directional aspects of deposystems from paleocurrent data and lithologic trends.

3.4.1 Proximal Basin Margin Deposits

Along the southern margin of the coalfield, fluvial sands and gravels of assemblage I and finer, poorly sorted pebbly sands and muds of assemblage II were deposited. The provenance of polymictic, predominantly granitic clasts, the northerly (basinward) decrease in maximum grain size (Blissenbach, 1952; Bluck, 1964; Boothroyd, 1972; Kesel and Lowe, 1987) and overall thinning of the deposits of these two assemblages point to a source area in the Cobequid Highland massif along the southern border of the basin. The trends in grain size and thickness coupled with scarce paleocurrent data for these deposits in the area of the Springhill coalfield indicates a predominantly northwesterly paleoflow transverse to the basin margin.

As discussed in the previous chapter, the basal conglomerate assemblage could be attributed to proximal to distal braided rivers and alluvial plains, transitional between assemblages G II and G III of Rust (1978). The essentially geomorphic question of whether these deposits formed alluvial fans or coalesced braidplains was deferred until this point.

On strictly sedimentological grounds, resolution between braidplain and distal alluvial fan deposits is difficult (Darby *et al.*, 1990). The predominance of waterlain deposits in the examined section on Polly Brook cannot be considered conclusive evidence of a braidplain. Both arid-zone and humid-zone alluvial fans, as well as braidplains, can be relatively rich in water-laid gravel deposits (Bluck, 1964; Hooke, 1967; Bull, 1977; Schumm, 1977; Kochel and Johnson, 1984). Humid, tropical alluvial fans of Costa Rica, adjacent the Cordillera de Talamanca, are constructed predominantly of braided river deposits, especially longitudinal braid bars (Kesel and Lowe, 1987). Even on geomorphic grounds, the distinction between coalesced alluvial fans and braidplains can be indistinct. Alluvial fans derived from the Panamint Range along the west side of Death Valley, for example, coalesce to form a piedmont essentially devoid of intra-fan areas.

The sedimentary processes inferred to have formed the deposits of lithofacies assemblage II are strongly indicative of low-gradient, distal fan deposition. Table 3.3 compares the morphologic, hydrologic, sedimentologic and botanical characteristics of the proximal basin-margin deposits of the Cumberland Basin at Springhill with those of modern humid-tropical to arid-zone alluvial fans. In the past there had been a striking disparity between the literature on humid-zone tropical and arid-zone fans, the latter being much more extensively documented, however several papers on tropical fans have been recently published (Table 3.3). Although the inland fans of the Kosi and Markanda rivers in northern India have been included as humid-tropical fans (Kochel and Johnson 1987), these "inland deltas" or "terminal fans" are essentially non-basin margin landforms (plains fans of North *et al.*, 1989) dominated by fluvial-channel deposition. They differ from the basin-margin deposits in the minor contribution of channelled flow in comparison with assemblages I and II and the other examples cited. The depositional processes of arid and humid-zone fans may be similar in many cases, with appreciable differences perhaps found only in pedogenesis and in lapse of time between major depositional events.

The dominant modes of deposition within assemblage II have been interpreted as sheetflow, poorly constrained channel flow and possible dilute mudflow. Local, discontinuous channels which locally collect sheetflow on low-gradient regions of the Dead Mesquite fan of Arizona (Packard, 1974, in Bull, 1977) and distal fans of Death Valley (Plate 2.4) is envisaged as the mode of deposition for the crudely stratified and ill-sorted pebbly/granular mudrocks and sandstones of assemblage II, as discussed in Chapter II. The diagnostic granular mudrocks (Fg) of this assemblage are strikingly similar to distal fan deposits of the Roaring River fan, Colorado pictured in Blair, (1987a) and similarly interpreted as having been deposited during deceleration of expanding sheetfloods.

The low gradient of the basin-margin deposits is supported not only by the occurrence of expanding sheetflows, but by the dominant mode of deposition. Hooke (1967) and Bull (1977) concluded that abundant water-laid deposits with possible subordinate debris-flow deposits were indicative of alluvial fans of gentle slope. This is congruent with the interpretation of the Springhill deposits. With progressively steeper slope, debris-flow and sieve deposits become dominant.

CRITERION	BASIN MARGIN DEPOSITS SPRINGHILL COALFIELD	HUMID-ZONE FANS			SEMI-ARID / ARID-ZONE FANS		
		N.E. PAPUA[1]	KOSI[2,3]	ROARING R.[4]	MARKANDA[5,6+]	DEAD MESQUITE[7]	S. NEVADA[8]
Fan Morphology							
Area: (Indiv.)	>40 Km^2 (assoc. II)	45 - 100 Km^2	16,500 $Km^{2,3}$	0.25 Km^2	38^5 - 64^6 Km^2	0.2 Km^2	12 Km^2
(Coalesc.)	>200 Km^2	100's Km^2	-	-	-	-	-
Thickness:	up to 100's m	-	900 m	14 m	-	-	-
Axial profile:	shallow gradient	shallow to steep?	nearly flat	steep	nearly flat	moderate	steep
Hydrology							
Annual precipitation:	(humid?)	up to 3000 m	1524 mm	-	540 mm	250 mm	?
Discharge:	flashy, ephemeral	monsoonal	flashy, monsoonal	catastrophic flood	ephemeral, monsoonal	flashy?	?ephemeral
Flow:	sheetflow, poorly constrained channel flow, (?dilute mudflow)	mudflow and/or streamflow	braided & meandering; sheetflow	debris flow, sheetflow, and subsequent braided channel flow	low sinuosity channel flow; sheetflood	sheetflow; poorly constrained braided channel flow	mudflow & subsequent channel flow
Triggering process:	rainstorm?	slumps (due to monsoon?)	rainfall	dam failure	monsoon	rainfall	-
Sedimentology							
Stratification:	crude horizontal, graded shallow scours, graded,	layered mudflow -	cross-strati- fication	crude horizontal to low angle	cross-stratification	(crude low-angle?)	crudely stratified
Major lithofacies:	pebbly mudrock (Fg) interstat. sandstone and mudstone (Si,Fi)	boulder clays (Gm); pebbly sandstone (Sg)	cross-stratified sands (St, Sp), mud drapes, inter- stratified sand and mud (Si, Fi)	pebbly mudrock (Fg) bouldering mud (Gm) pebbly sandstone (Sg)	cross-stratificated sands (St, Sp), horizontal strad'd sands (Sh); interstrat. sand and mud (Si, Fi)	"sands and silts"	pebble & boulder conglomerate (Gm) sands?
Sand body geometry:	single-storied	-	hollow fills, benches	-	single, rarely multi-storied. hollow-fills, barforms	broad swales, hollow-fills	hollow-fills
Pedogenesis							
Bioturbation (roots):	common (assoc. II)	-	common	-	common	common	-
Soils:	rare type 1 caliche	strongly weathered	gleysols, calcrete	-	calcrete	(desert varnish?)	strongly weathered
Vegetation							
Fan surface:	vegetated	(vegetated?)	vegetated	(vegetated?)	vegetated	vegetated	-
Alluvial apron:	peat mire wetland	freshwater swamps at toes of large fans	"flood basin marshes" in places	forested woodland	"swampy"	(desert?)	-

Sources: [1]Ruxton, 1970; [2]Gole and Chitale, 1966; [3]Wells and Dore, 1987; [4]Blair, 1987; [5]Mukherji, 1976; [6]Parkash et al., 1983; [7]Packard, 1974; [8]Bluck, 1964

Table 3.3 Summary characteristics of arid and humid-zone alluvial fans.

In summary, the basin-margin deposits of the Springhill coalfield formed an alluvial piedmont characterized by braided stream deposition and subsequent sheetflow to poorly constrained channel flow. The sedimentology, areal distribution and transverse flow adjacent the Cobequid massif are consistent with low gradient alluvial fans.

3.4.2 Basin-Floor Deposits

The deposits of the basin distal to the alluvial fan sheetflood deposits of the piedmont include lithofacies assemblages III, IV and VI.

Coal-bearing assemblage IV comprises three sub-assemblages: i) multistorey sandstone; ii) mudrock-dominated; and iii) coal sub-assemblages. Orientation of channel-belts within the multistorey sandstone sub-assemblage (Figure 3.8) is northeasterly, indicating that flow ranged from oblique to parallel to the basin margin. Other paleoflow indicators, however, show varied directions (Figure 3.7), possibly a reflection of a sinuous drainage system. It is not always possible, however, to deduce river planform by paleoflow variance (Jackson, 1978). Regional paleoflow within the Cumberland and Maritimes Basin was predominantly northeasterly (Calder *et al.*, 1988; Gibling *et al.*, in press), however, local flow reversal into the basin (cf. Stellarton Basin; Yeo and Ruixiang, 1987) should not be discounted without further data.

Forested peat mires flourished for millenia at sites basinward of the distal alluvial fan toes (Figures 3.6, 3.9). The lithology of siliciclastic partings along the southern margins (piedmont zone) of the coal seams shows that sheetfloods of fine, poorly sorted fan deposits periodically swept into the mires. The characteristically sharp southern boundary of the inner mire zone could have arisen at least in part from the baffling effect of dense forest swamp vegetation. Such a boundary also may indicate that the mires exhibited a southward dipping, marginal slope as in some modern tropical forest swamps (Anderson, 1964) and analogous to the "rand" (Gore, 1983) of modern temperate-zone raised bogs.

Basinward of the mires, major river systems occupied channel belts of 1 to 2 km width (Figures 3.6, 3.9). The northern margins (riverine zone) of the mires were inundated by major floods which deposited mud-rich sediment. The resulting mudrock partings commonly extend 500 to 1000 m into the seam, somewhat farther than the 400 m extent of swamp flooding adjacent to the Baram River of Brunei as reported by Anderson (1964).

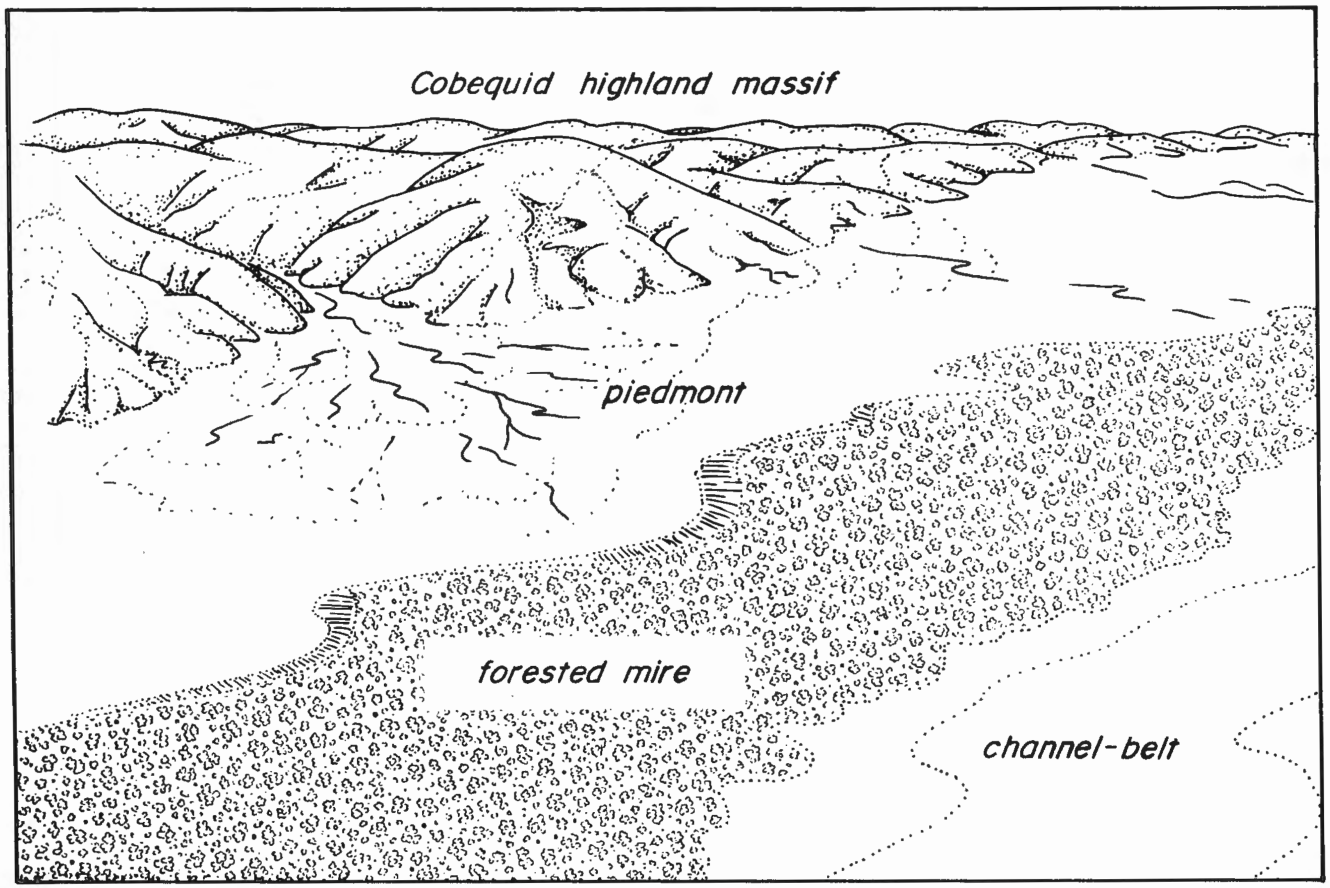


Figure 3.9 Paleogeographic reconstruction of the Springhill coalfield (No. 3 paleomire depicted).

The mires in the riverine zone were on occasion submerged to the point where peat ceased to accumulate and subaqueous sideritic muds were deposited. Long-term flooding of this sort was followed by avulsion of the river to a site overlying the submerged, mud-laden northern mire margins. Subsequent basinward (northward) avulsion was followed by re-establishment of the mire, resulting in the entombment of multistorey sandstone bodies such as those found within the No. 3, No. 1 and McCarthy seams.

The lithosome distribution map (Figure 3.6) depicts the areal position of sediment bodies/type during partial coverage of the No. 3 seam peat mire by the channel belt near mid-life of the mire. A 700 m wide peripheral zone of mud and silt between the mire and river belt is analogous to the "alluvial strip" between the Baram River and neighboring mires of Borneo which varies from 30 - 300 m wide (Wilford, 1962).

3.4.3 Temporal Changes in Basin-Filling

The stratigraphy and sedimentology of lithofacies assemblages show that the depositional environments varied as the basin was infilled. The reconstructed stratigraphic section of the coalfield (Figure 3.2) illustrates that the assemblages young progressively from III, IV, V, through VI. Referring to the summary of environmental interpretation for the individual lithofacies assemblages (Table 2.6), this younging sequence of assemblages indicates a change in the basin-floor environment from wetland-fringed lacustrine shoreline (assemblage III - IIIB) to a predominantly submerged alluvial plain traversed by rivers of variable discharge (assemblage IV). This environment was most conducive to lush forest mires which, in positions distal to the fan toes of the basin margin, may have accumulated up to 20-40 m of peat (see Constraints in the Interpretation of Time, Chapter V).

The areal zone in which perennially reducing conditions prevailed and thick peats accumulated became progressively diminished as basin-filling proceeded. This trend is reflected in the restriction of areas of major peat accumulation in seams above the No. 2 (Figures 3.2 and 3.5). Areas marginal to these shrinking paludal districts were sites of fluctuating ground water-level, subaerial exposure and gleying of muddy soils (Bown and Kraus, 1981) reflected by mottled deposits of assemblage VI. With time, groundwater levels became increasingly lower as the virtually wholesale reddening of mudrock in lithofacies assemblage VII attest. At this late stage of basin filling, saturated areas were restricted only to topographic depressions within the confines of the channel belt where grey, sideritic mudrock lenses occur. The accumulation of peat by this time was virtually impossible due to the high levels of oxidation. The question of whether such changes were due primarily to climatic or paleogeographic/drainage factors is addressed later.

3.5 WETLANDS ASSOCIATED WITH ALLUVIAL FANS: POSSIBLE ANALOGUES

In both arid and humid-zone settings, wetlands commonly occur on the alluvial apron of alluvial fans. This relationship has gone virtually unrecognized when considering models of ancient peat (coal) formation, notable exceptions being Galloway and Hobday (1983) and Flores (1986). In the main, the proximity of alluvial fan deposits has been considered a detriment to peat/coal formation. The acceptance of a positive influence exerted by such depositional systems requires a re-appraisal of theoretical conditions for "coal deposition" (cf. McCabe, 1984).

Alluvial fan deposits, particularly those with a mud-poor framework, are highly transmissive aquifers. Distal areas of alluvial fans, where topographic and hydraulic gradients decrease markedly, are zones of major shallow meteoric groundwater discharge (Galloway and Hobday, 1983). The alluvial apron is thus characterized by a shallow to emergent water table, with ion-rich waters that tend to be reduced and of elevated pH (Figure 3.10). These conditions promote plant growth and preservation of organic accumulation. The ponding of groundwater within the distal fans by muddy deposits at the fan toes can result in the formation of springs along the distal fans, as in Death Valley (Hunt, 1975). Extensive alluvial fans along the west side of Death Valley discharge an estimated 2,850 gallons (US) per minute from a number of springs and seeps whereas the steep, short fans on the east side adjacent to Badwater yield only 70 gals/min. (Hunt, 1975). The difference has been attributed by Hunt to the greater height and area of the Panamint Mountains on the west side.

The importance of alluvial fan-derived meteoric groundwater discharge in the support of wetland vegetation is most apparent in regions of low rainfall. Alluvial fan-associated wetlands occur along the axis of the Great Valley of California (Galloway and Hobday, 1983), and buried peaty beds have been intersected in groundwater drilling (W. Galloway, personal communication, 1991). Even in the extreme aridity of Death Valley, vegetation occurs along distal fan toes (Plate 3.1). At the foot of alluvial fans, particularly those that are extensive, groundwater is ponded against the side of muddy sediments. The resulting shallow groundwater, seeps and springs support a well defined belt of phreatophytes (plants whose roots extend to the water table), notably Mesquite (Hunt, 1975; Plate 3.1). Certain spring-fed wetlands may accumulate peat, an example being the macaques of Moçambique (Thompson and Hamilton, 1983). In northern India, the semi-arid Markanda plains fan is described as "swampy", with a near surface water table (Mukherji, 1976).

RECHARGE		DISCHARGE
--------------------	pH	------------------>
<-----------------	Eh	------------------
<-----------------	Ca^{++}	------------------
-------------------	Na^{+}	------------------>
-------------------	Cl^{-}	------------------>
Variable ------------------	$SO_4^{=}$	------------- H_2O, HS^{-}
-------------------	HCO_3^{-}	------------------>
<-----------------	Si	------------------

(--------> increasing)

Figure 3.10 General downflow trends in hydrochemical evolution of groundwater, after Galloway and Hobday (1983).

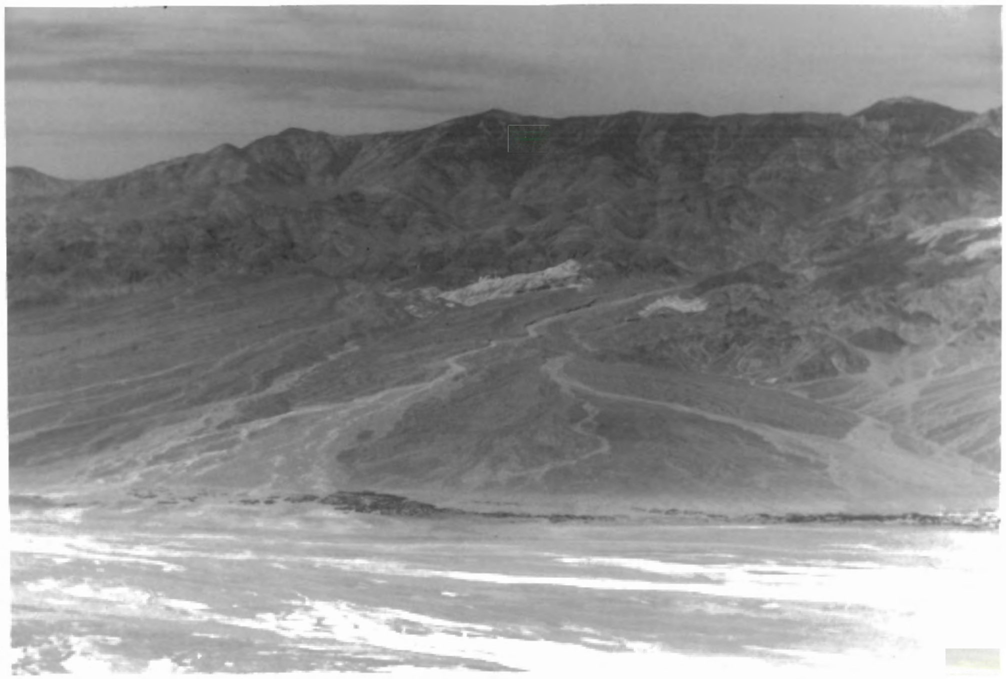


Plate 3.1 a) Starvation Canyon alluvial fan, emanating from the Panamint Mts., west side of Death Valley. Dark area fringing foot of fan is zone of vegetation (mesquite and other phreatophytes) fed by groundwater discharge.

b) Mesquite (indigenous) and date palms (introduced) near Furnace Creek, Death Valley, fed by groundwater springs at the toes of coalesced alluvial fans derived from Funeral Mts., in background. Springs that feed this particular stand of vegetation may be sourced in part in mountains to the east of the basin.

Peatlands in a humid temperate setting occur adjacent to alluvial fans associated with the strike-slip Alpine Fault system on South Island, New Zealand (J. S. Esterle, written communication, 1990; J. Shearer, written communication, 1990). In humid tropical northeast Papua New Guinea, freshwater swamps occur distal to alluvial fans (Ruxton, 1970). Although Ruxton did not state whether these wetlands were peat-forming, much of the rainforest accumulates peat (Anderson, 1983). Alluvial fans border extensive peat-forming rainforest of the humid tropical Markanda Basin, southern Kalimantan, Indonesia (Flores, 1986).

3.6 STRUCTURAL CONTROLS ON CHANNEL-BELT AND MIRE LOCATION

The hypothesis that structure and especially faults active during deposition played a role in controlling peat-forming and fluvial systems is by no means novel. Fault control of river course and peatland location has been proposed for Carboniferous coalfields of the Appalachian Basin, U.S.A. (Horne *et al.*, 1978; Ferm and Staub, 1984; Weisenfluh and Ferm, 1984) and for intermontane coalfields of France and Spain (Courel *et al.*, 1986). Although such a proposed control may seem simplistic, one is hard pressed to ignore the apparently similar trends of major faults, multistorey sandstone bodies (Figure 3.8) and zones of maximum coal accumulation (Figure 3.5) in the Springhill coalfield. One is faced with four possible explanations: i) the similarity of trends is an unrelated coincidence, ii) faults played a role in channel-belt and mire location, iii) faults are a consequence of the arrangement of multistorey sandstone bodies and coal seams, perhaps related to differential compaction, or iv) local faults, sandstone body and coal seam orientation were all determined by a higher order control such as regional basin configuration, but were independent of one another.

The similar trends of the southern boundaries of multistorey sandstone bodies within riverine zone partings of the Nos 1 and 3 seams, of the Rodney Sandstone body, and of major faults are striking (Figure 3.8). While not conclusive, the similar trends point to the possibility that faults may have been active during deposition and to some degree influenced local topography, hence channel-belt course, groundwater level and mire development.

3.7 CORRELATION OF COAL-BEARING STRATA WITHIN THE CUMBERLAND BASIN

The stratigraphic correlation of the Springhill and Joggins coal-bearing sections has long been a contentious issue, the only point of agreement being that the Springhill coals are *no older* than those of Joggins and that the two sections are in part correlative. Bell (1938) placed the Springhill coals somewhat higher in the stratigraphic section than the Joggins coals. He considered the bivalve-associated coals of the Salt Springs area to lie near the horizon of the Joggins seam and below the No. 6 seam of Springhill. Copeland (1959) generally concurred with this appraisal. Shaw (1951), largely on the basis of structural relationships traceable in the field, placed the composite Chignecto seam (Kimberly-Forty Brine equivalent, Bell, 1938; Copeland, 1959) at the horizon of the No. 2 seam of the Springhill section. Hacquebard and Donaldson (1964) concluded that the Springhill and Joggins sections were stratigraphically equivalent on the basis of calculated depth of burial utilizing rank of the coals (Suggate index) and on the basis of similar spore assemblages, in particular *Lycospora* abundance peaks. The use of the Suggate index for correlation assumes that subsequent basin filling was symmetrical.

At Springhill, strata assigned to the Joggins Formation (lithofacies assemblage III) occur stratigraphically below the Gesner seam. The contact between the Springhill Mines and Joggins formations, however, is apparently diachronous and it should not be assumed on this basis that all coal seams within the Springhill Mines Formation at Springhill lie stratigraphically above those of the Joggins Formation exposed on the Joggins shore.

3.7.1 Seismic Correlation

Recent high-resolution seismic data acquired by the Geological Survey of Canada (Bromley, 1987) and interpreted by the writer and by D. S. Bromley of the GSC (Calder and Bromley, 1988; in press) suggest a correlation of Joggins and Springhill seams similar to that of Bell (1938). Reflecting events were correlated with individual coal seams of the Springhill coalfield with the aid of drillholes, or projection to mine workings and/or known subcrop positions. Reflectors at the horizons of the No. 6 and No. 2 seam were traced from Springhill, west-northwesterly to Athol and thence north-northwest to Maccan. Two vertical zones of poor response near the tie-in at Athol present possible sources of error in correlation; however, the similar character of reflectors adjacent to these zones would appear to allow only minor correlative misalignment.

The reflector/horizon of the No. 6 seam, when projected to outcrop at Maccan, occurs at or near the position of the Joggins seam (coal 7 of Division IV of Logan, 1845). The reflector at the horizon of the No. 2 seam projects approximately 360m above the Joggins seam. In projecting this horizon westward, the No. 2 seam horizon is inferred to occur at or near the position of several thin seams that outcrop immediately south of MacCarron's Creek (coals 15 to 19 of Logan's Division III).

Such a correlation places the upper, major coal seams (Rodney, McCarthy, Nos. 3, 1 and 2) of the Springhill coalfield in the lower half of Logan's Division III near the horizons of several thin, poorly developed coals (9-19). The lower, major coals seams of Springhill (Nos. 7 and 6) lie near the position of the highest coal historically deemed mineable at Joggins (coal 7, Division IV of Logan, 1845, also known as the Joggins seam).

The tie-in through seismic profiles of Springhill seam horizons with seams of the Joggins-Chignecto coalfield at Maccan places the position of the Gesner seam near that of the Forty Brine/Chignecto seam. Such a correlation supports approximate correlation of the Gesner, Sandrun and Forty Brine/Chignecto seams on the basis of similar associations of lithofacies (Calder and Naylor, 1985) and position in the vertical basin-fill sequences at Springhill, Salt Springs and Joggins.

The hypothesis that strikingly well defined ("bright") reflecting events within the basin-fill correlate with strata comprising multistorey sandstone bodies and thick mudrocks and potentially coal seams ('cyclothems') as proposed by Calder and Bromley (1988) was tested and proven by the drilling of the 1156 m corehole SA.88.2 at South Athol (Figure 3.11). The sequence of thin (<1 m) coals intersected between 1090 and 1152 m are believed to correlate approximately with the upper, thick coals (No. 3 - Rodney seam) of the Springhill coalfield (Calder and Bromley, in press). Such seismic signatures offer promise as exploration targets, but do not indicate unequivocally that coals are present.

3.7.1.1 Other coal-bearing districts of the Cumberland Basin:

Other coal-bearing districts within the Cumberland Basin are structurally dislocated from the Joggins-Springhill coalfields of the Athol Syncline. These notable occurrences of coal-bearing strata are i) coals intersected in the Pettigrew borehole near Newville (Halfway) Lake; ii) coals of the Salt Springs district; and iii) coals west of the Athol Syncline at Roslin.

The Pettigrew borehole (Figure 1.3), also known as Standard Coal and Railway Co. borehole No. 1, reportedly intersected 2.59 m of coal at a depth of 716.46-719.05 m (Brown, 1907). The authenticity of the reported intersection however, has long been rumoured to be questionable (Copeland, 1959). Bell (1938) believed the Pettigrew seam to be stratigraphically equivalent to the upper coals of Springhill, a correlation supported by the writer. The difficulty in correlating the Pettigrew seam with those of Joggins and Springhill arises i) from the fact that the coal-bearing strata intersected in the Pettigrew borehole apparently do not outcrop, and ii) it is now known that the

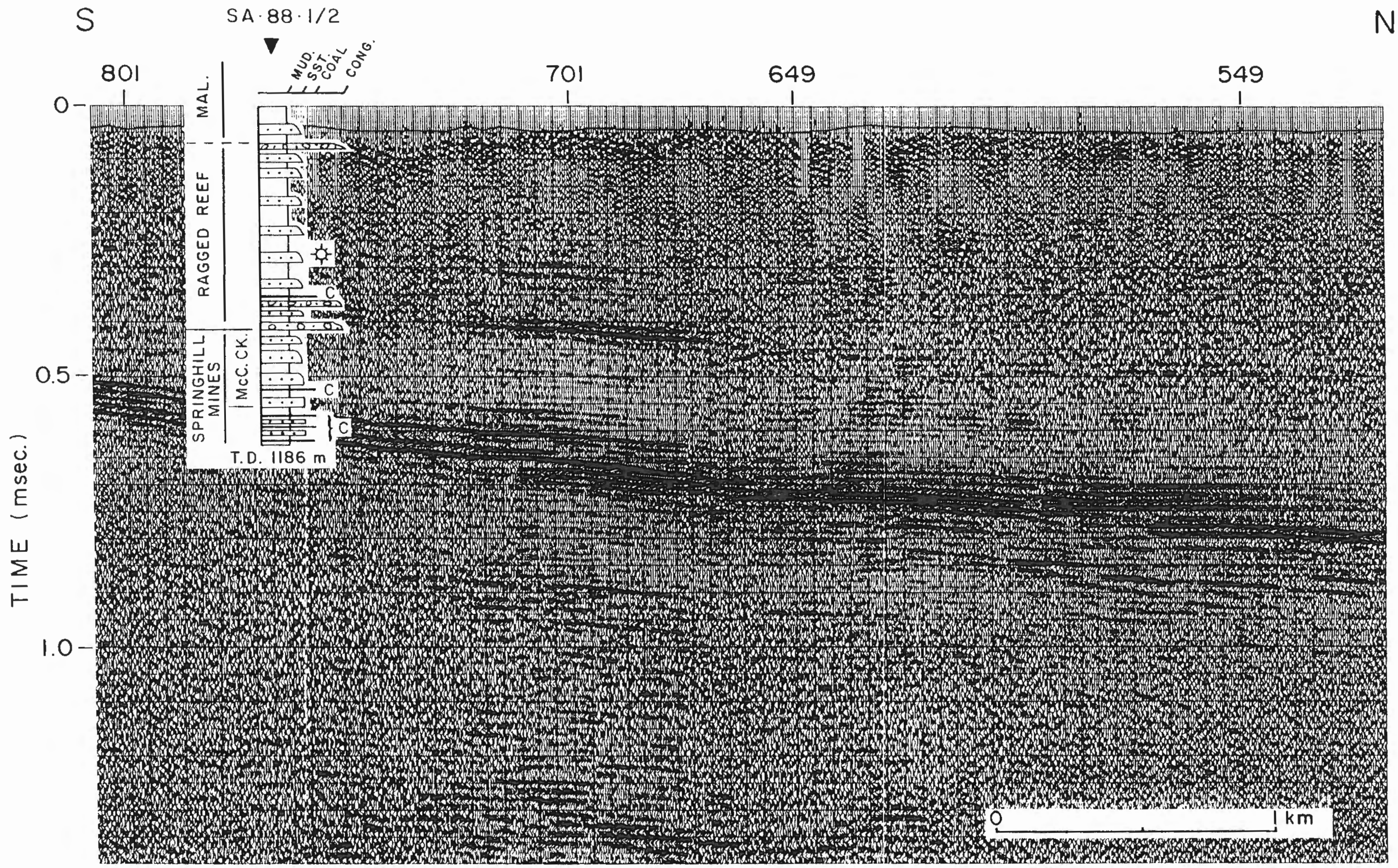


Figure 3.11 Seismic signature of cyclic multistorey sandstone and mudrock proven here to be coal-bearing by the drilling of the South Athol drillhole SA.88.2.

Pettigrew seam is dislocated from the Joggins and Springhill coals by the Athol-Sand River fault zone (Calder and Bromley, 1988; Ryan *et al.*, 1990). Correlation is further hampered by the absence of core and vague description of strata (Brown, 1907) from the Pettigrew borehole.

Although the correlation is far from certain, the GSC seismic data appear to support the belief of Bell (1938) that the Pettigrew seam is equivalent to one of the upper Springhill coals. This tentative correlation is based on the stratigraphic position of the Pettigrew seam above conglomerates inferred to be of the Polly Brook Formation: an onlap relationship may therefore exist here, as with the Rodney, McCarthy and No. 3 seams of the Springhill coalfield.

Coals of the Salt Springs district occur within strata that outcrop on Black River northeast of Springhill. These northerly dipping strata are structurally dislocated from the Springhill coalfield to the southwest and the Joggins-Chignecto coalfield to the north by diapiric Windsor Group strata which form the Black River diapir at the western terminus of the Claremont Anticline. The Salt Springs coals are inferred to occur near the top of the Joggins Formation in this area on the basis of their association with pelecypod-bearing strata and by their position overlying polymictic conglomerates tentatively assigned to the Polly Brook Formation. The Salt Springs coals therefore may occur at a stratigraphic position near the lowermost Springhill coals such as the Gesner seam, which lies near the boundary of the Springhill Mines and Joggins Formation.

Coals west of the Athol Syncline include the Jungle seam, south of Oxford Junction on Jungle Road, and the coals of the Roslin district. The Roslin coals are overlain by thick pelecypod-bearing mudrocks and underlain by extrabasinal polymictic conglomerates and intercalated bivalve-bearing mudrock (Calder and Naylor, 1985). These strata, covered by a thick mantle of Pleistocene deposits, were cored in drillholes OX-81-9 and R-1. Their correlation with coals of the Athol Syncline is uncertain. The coal-bearing strata at Roslin are considered by the author to belong to the Joggins Formation on the basis of lithology, in particular the presence of bivalve-bearing mudrock (Calder and Naylor 1985). This interpretation has been subsequently adopted by Ryan *et al.* (1990).

3.7.2 Biostratigraphic Correlation with Springhill

3.7.2.1 Joggins

The detailed palynological studies of coal-bearing strata in the Cumberland Basin by Dolby (1984, 1986, 1988a, 1988b, 1991) have permitted the palynostratigraphic correlation of coal measures (Figure 3.12), particularly between the Springhill and Joggins sections. The two sections are particularly difficult to correlate because there are few species extinctions or appearances in the late Westphalian A - early Westphalian B in the European Standard Sequence (Dolby, 1991) and because extinctions and appearances in the Athol Syncline tend to be drawn out across the great thickness of basin-fill. Fifteen palynological events deemed to be stratigraphically significant were plotted by Dolby relative to the base of the Joggins and Springhill Mines Formation, and best-fit correlation line drawn (Figure 3.13). Two points that show a relatively weak correlation are the base at Springhill of *Vestispora tortuosa*, which is taken from corehole E89-01, of questionable correlation, and the base of *Cananoropollis mehtae*, which may have been previously mis-identified and may be present lower in the Joggins section (Dolby, 1991).

The correlation (Figures 3.12, 3.13) indicates that the best developed seams at Springhill lie stratigraphically above those at Joggins, as proposed by Bell (1938). The Gesner seam at Springhill and the Queens/Joggins seams at Joggins are thought to be time-equivalent (Dolby, 1991). The No. 2 seam is inferred to be coeval with thin coals exposed near MacCarrons Creek. The biostratigraphic correlation of Dolby (ibid.) is in general agreement with the seismic correlation of the Springhill and Joggins sections (Calder and Bromley, 1988; in press).

3.7.2.2 South Athol

The coals intersected at depth in the South Athol corehole SA.88.2 (Calder and Gillis, 1989) are believed to be no older than zone 4 of Dolby (1991) on the basis of *Vestispora tortuosa* found at 1125 m. The occurrence of *Lophotriletes ibrahimii* and *L. "spinosaetosus"* in the Epsilon seam at 1152 m correlates with sample MB-TR-30 (see Ryan *et al.*, 1990) in the MacCarrens Creek coals of the Joggins section (Figure 3.12).

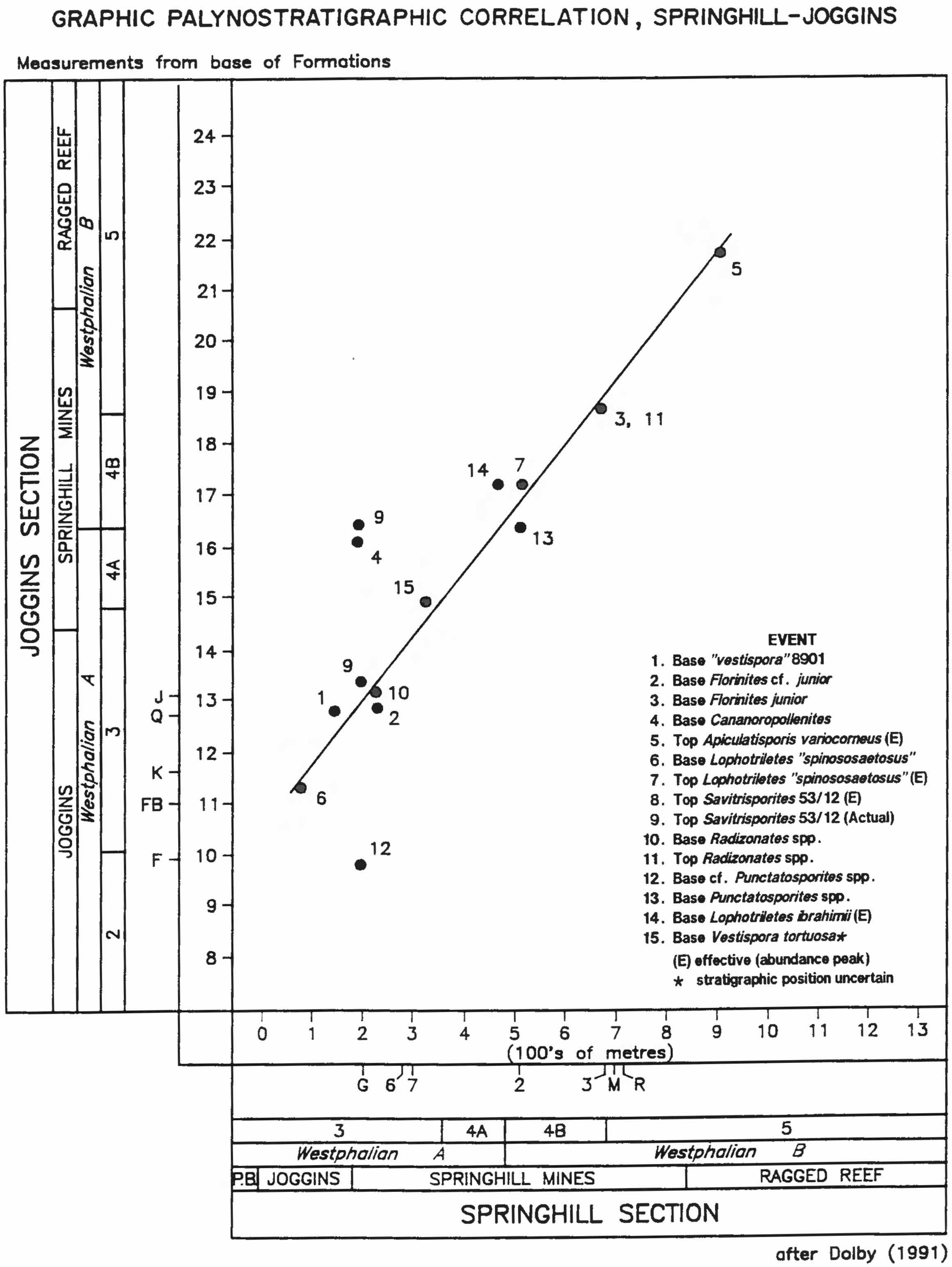


Figure 3.12 Graphic palynostratigraphic correlation of the Joggins and Springhill coal-bearing sections (after Dolby, 1991).

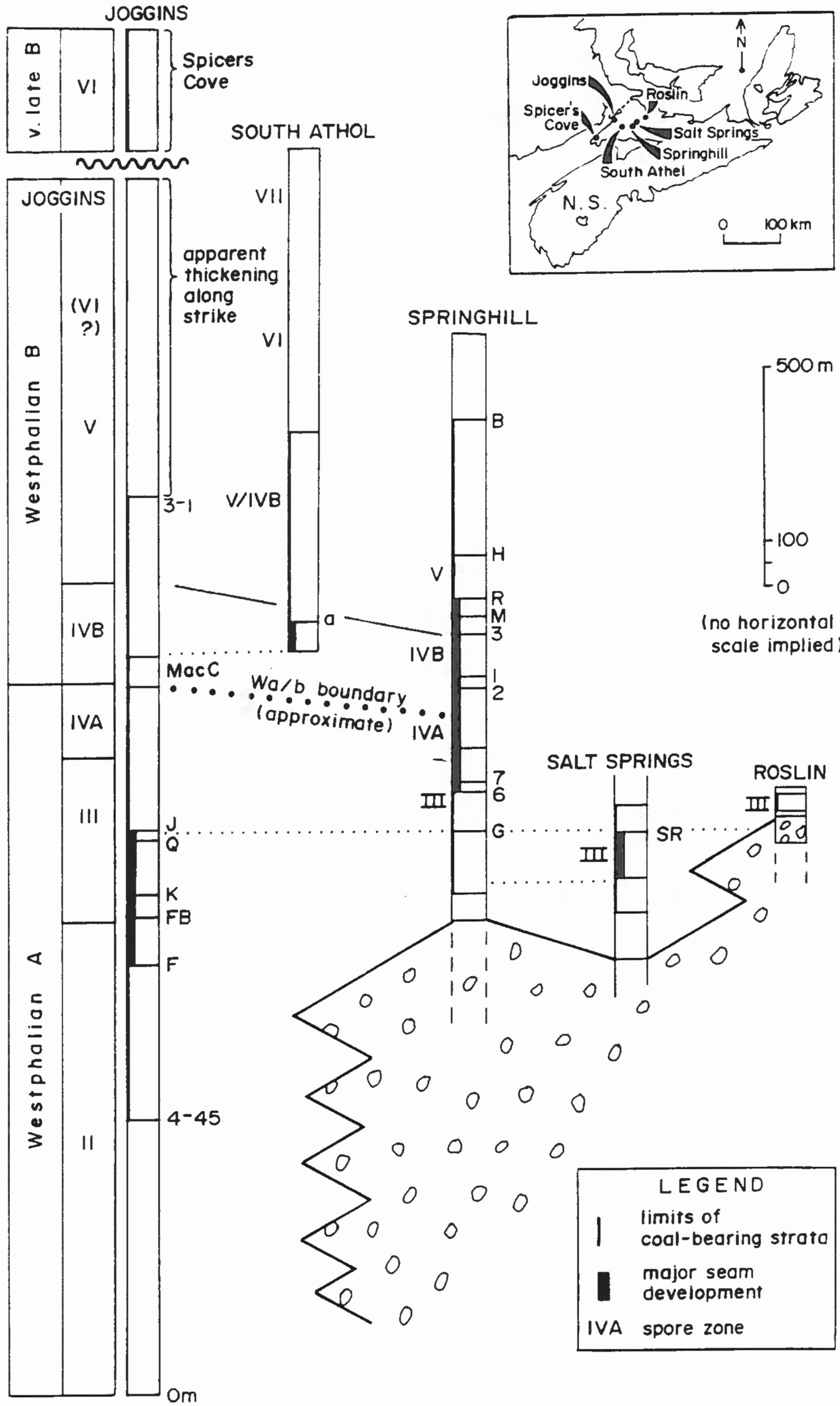


Figure 3.13 Ages of coal-bearing strata in the western and central Cumberland Basin, based on miospore zonation of Dolby (1991). Coal seam nomenclature: Joggins: F-Fundy, FB-Forty Brine, K-Kimberly, Q-Queen, J-Joggins, Mac-MacCarrons River seams; Springhill: G-Gesner, M-McCarthy, R-Rodney, H-Harrison, B-Barlow; Salt Springs: SR-Sandrun.

3.7.2.3 Salt Springs and Roslin

Fifteen samples from the Salt Springs coal-bearing sequence spanning a stratigraphic interval of 285 m, were submitted for palynological analysis. The sequence was tentatively assigned to zone 3, and the Sandrun seam (SR; Figure 3.12) inferred to be approximately contemporaneous with the Gesner seam of Springhill and the Joggins-Queen seams of Joggins (Dolby, 1991).

Coal-bearing strata at Roslin (Figure 1.3) comprising mudrocks bearing an abundant bivalve fauna and coal, intercalated with, and overlying polymictic conglomerates (Calder and Naylor, 1985) are deemed to be of similar age to the Salt Springs coal sequence (Dolby, 1991).

The Salt Springs and Roslin sections were correlated with the Springhill section (Dolby, 1991) largely on the basis of the first occurrence of *Punctatosporites spp*. and *Cananoropollis mehtae* and the last occurrence of *Savritrisporites* 53/12 (provisional designation of Dolby), which coincide in the Gesner seam of the Springhill coal-bearing sequence. The similar age of these pelecypod-bearing strata supports the lithostratigraphic correlation of Calder and Naylor (1985).

3.8 SEQUENCES AND CYCLES WITHIN THE BASIN-FILL AT SPRINGHILL

Several orders of cyclicity are apparent within the basin-fill record in the Springhill coalfield (Fig. 3.14) ranging from metre to kilometre - scale and spanning a wide temporal range (discussed in the concluding chapter). Cyclicity of coal-bearing strata has been long-recognized, some of the earliest observations being made from the classic Joggins section of the Cumberland Basin (Dawson, 1855, among others). The best known and most widely applied (and indeed perhaps least understood) term with respect to a coal-bearing cycle is *cyclothem*, broadly defined by Wanless and Weller (1932, p. 1003) as "a series of beds deposited during a single sedimentary cycle of the type that prevailed during the Pennsylvanian". The ambiguity of the term has led to a wide range of application (see Duff *et al.* 1967 for a comprehensive discussion).

Cycles within the basin-fill of the Springhill coalfield (Figure 3.14) include in order of decreasing magnitude: 1) basin-fill sequence in the order of 1000 m thick; 2) intervening sequences marked by the occurrence of red mudrocks apparently correlative (Figure 3.2) with basin-margin conglomerate tongues, the thickness between red bed-fan events being in the order of 100's of meters; 3) nonmarine "cyclothems" (sensu latu) comprising the succession of coal, sideritic mudrock locally

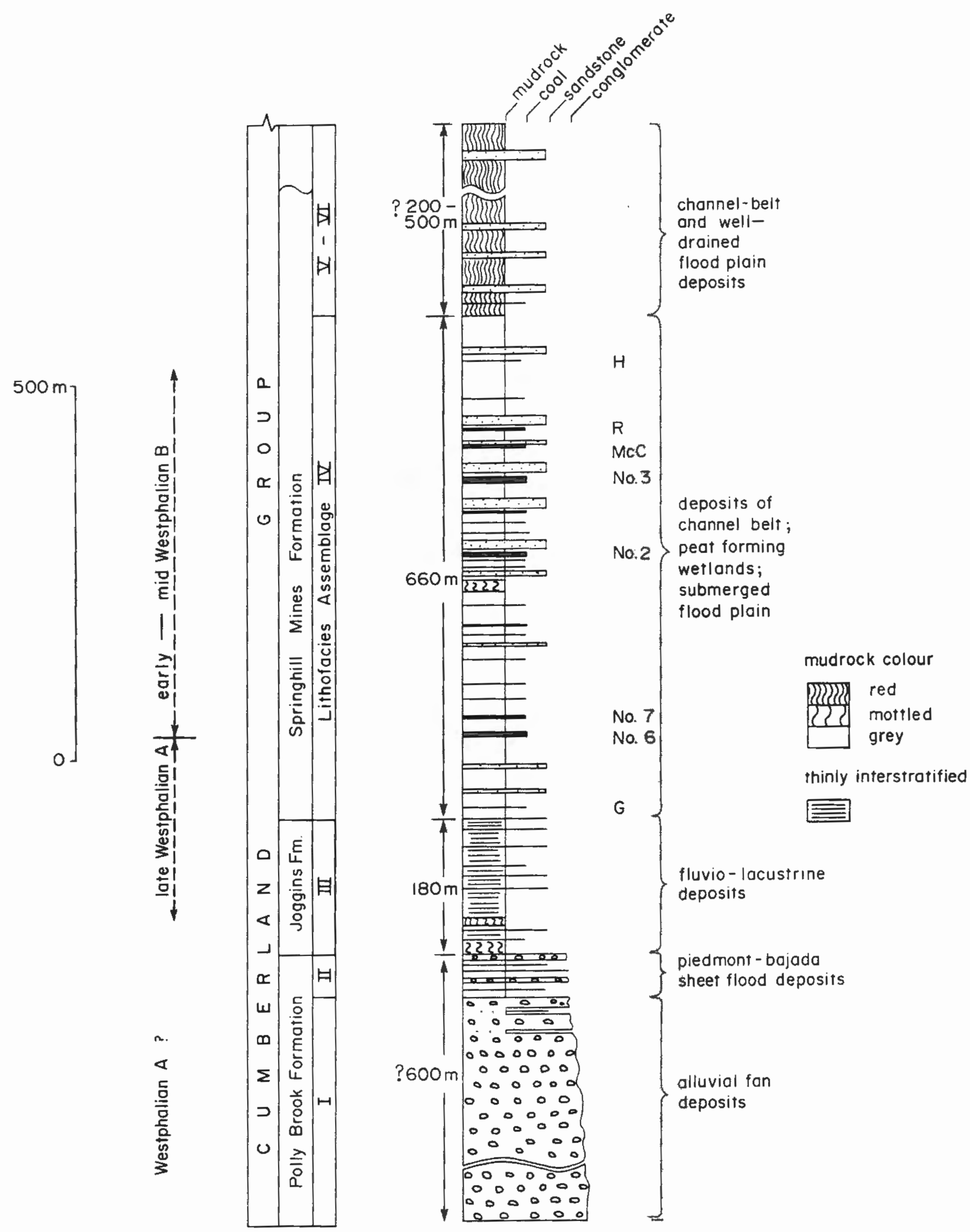


Figure 3.14 The basin-fill sequence represented by the Cumberland Group in the Springhill coalfield. The Rodney seam and overlying sandstone body comprise part of the cyclothem depicted in Fig. 3.15. Coal seams:G-Gesner, R-Rodney, H-Harrison.

with cone-in-cone limestone, multistorey sandstone or laterally associated sheet sandstones, coal, together comprising a total thickness in the order of 20-30 m; and 4) cycles of intraseam scale, including those of coal lithotype and seam partings, dm to m thick. Still smaller-scale cycles of dm- to mm-scale occur within siliciclastic rocks and of mm- to micron-scale within coal, the latter requiring microscopic examination of polished blocks.

3.8.1 Basin-Fill Sequence:

The Westphalian basin-fill sequence at Springhill (Figure 3.14), greater than 2 km in thickness, fines and ultimately reddens upward from the basal polymictic pebble-cobble orthoconglomerate and pebbly mudrock of assemblages I-II (Polly Brook Formation), inferred to be in the order of 600 m thick. The early coal-bearing portion of the basin-fill sequence is represented by <200 m of thinly interstratified mudrocks and sandstones and rare, thin coals of assemblage III (Joggins Formation). Overlying these strata are *ca.* 600 m of thick coals (<4.3 m), grey mudrocks and multistorey sandstone bodies (<20 m) of assemblage IV (the coal measures of the Springhill Mines Formation). Mudrocks become progressively redder and coal seams decline in abundance within the upper strata of the basin-fill sequence, the transitional assemblage V (MacCarrens River Member, Springhill Mines Formation) and assemblage VI (unassigned beds that underlie or are laterally transitional to conglomerates of the Ragged Reef Formation.)

A similar fining and upward-reddening basin-fill sequence is common to many coal basins, particularly of intermontane setting. Carboniferous coal basins exhibiting similar sequence within their basin-fill include the Stellarton Basin (Yeo and Ruixiang, 1987; Naylor *et al.*, 1989), and Debert-Kemptown Basin (Lortie, 1979; Calder, 1985) within the Maritimes Basin. A Mesozoic example is the Fuxin Basin of China (Li Sitian *et al.*, 1984).

The basin-fill sequence at Springhill records progressive evolution of increasingly distal alluvial fans, followed by lakes and subsequently by throughflowing rivers, perennially submerged floodplains and piedmont mires. Continued infilling of the basin witnessed better drainage, subaerial exposure of floodplains and diminished peatlands. Such a history of basin-filling may be primarily tectonic-driven, climatically-driven, or both. The ultimate deposition of red beds within the basin-fill has been attributed to increasing aridity (Rust *et al.*, 1984). If long-term *widespread* climatic change to more arid conditions were responsible for the basin-fill sequence then a chronostratigraphic correlation of redbeds within the Maritimes Basin should be apparent, but such is not the case (the

Stephanian-Permian red beds are a notable exception). Indeed, red beds are diachronous even within the Athol Syncline of the Cumberland Basin. Climatic change for the origin of red bed sequences in the Pennine Basin, U.K., was discounted on similar grounds (Besly, 1988). Local climate change induced by rising and subsequent denudation of the Cobequid Highland massif (cf. Perlmutter and Matthews, 1989) could, however, have been a factor in the evolution of the basin-fill sequence. Paleoclimate is discussed further in the concluding chapter.

The basin-fill sequence is interpreted to represent an upward change from an active basin margin to one of declining activity. Similar sequences have been reported from the Torridonian of Scotland and explained by retreat of basin-margin fault scarps (Bluck, 1967; Williams 1969). The upward change from medial to distal fan deposits and progressive basin margin onlap of lithofacies assemblages and coal seams of the Springhill coalfield (Figure 3.2) likewise support declining subsidence, basin margin relief and fault activity.

3.8.2 Red Bed-Conglomerate Events:

The basin-fill sequence is interrupted sporadically by red mudrocks in the order of a few 10's of metres thick. Red mudrocks recur at 100-400 m intervals; at the base of assemblage II (Joggins Formation), 30 m below the No. 2 seam, between the No. 1 and No. 3 seams in the north of the coalfield, and between the Harrison and Rodney seams. Examination of the north-south stratigraphic section (Figures 3.1, 3.2) reveals that these zones of red mudrocks occur at similar stratigraphic positions to tongues of pebbly mudrock and other poorly sorted sediments of assemblage II (Leamington Member, Polly Brook Formation).

The spatial connection between these two lithologies indicates a genetic relationship between alluvial fan progradation and gleying of floodplain muds under relatively low groundwater levels. Such a relationship between alluvial fan progradation and low periods of groundwater level appears to support the theory advanced by Blair (1987) and Blair and Biloudeau (1988) that fan progradation is favoured by periods of relatively slow subsidence. It appears likely, therefore, that these episodes of red mudrock formation and correlative alluvial fan progradation are of tectonic origin.

3.8.3 Coal-Mudrock-Sandstone 'Cyclothems':

Cyclicity within the basin-fill sequence is perhaps most apparent in the sequence of coal/sideritic mudrock, locally with cone-in-cone limestone/multistorey sandstone or associated interstratified sandstone and mudrock/rooted mudrock/coal. Such a sequence can be considered a variety of the non-marine cyclothem (Weller, 1961). One such cyclothem from the Springhill coalfield includes the Rodney seam and Rodney Sandstone (Figure 3.15).

Cyclothems that contain the past-mined seams of the Springhill coalfield range in thickness from 11-38 m, with a mean of 24 m. The inevitable association of thick coal and overlying multistorey sandstone bodies was a causative factor in the occurrence of rock bursts (bumps) associated with underground mining in the Springhill coalfield (Rice, 1924; Notley, 1984).

Springhill cyclothems represent the deposits of piedmont mires and throughflowing, basin axis rivers and associated floodplains. The most important question regarding these systems is whether they coexisted side-by-side as suggested by Walther's Law of Facies, or whether they represent temporally distinct environments which responded to episodic, allocyclic controls (Anderson and Goodwin, 1990). Stratigraphic relationships suggest that Walther's Law is, in this case, applicable, but with reservations: possible climate change during cyclothem development (see Chapter V) would have exerted an influence both on mire and fluvial development.

The processes by which the life of a mire was ended, and burial of the mire by floodplain and subsequently by channel-belt deposits is one of the most fundamental yet poorly understood in coal geology. One must search for a control or process which occurred with a pronounced regularity of time and intensity giving rise to cyclothems of a characteristic thickness within a given basin. The process may have been autocyclic, allocyclic, or both. Possible autogenic processes include: i) natural channel belt process such as avulsion, or ii) an intrinsic limitation to mire growth and peat accumulation. Possible allogenic processes include: i) eustatic change driven by climate (Wanless and Shepard, 1936; Heckel, 1986, Cobb *et al.*, 1989, among others) by tectonism (Klein and Willard, 1988) or by geoidal change (Mörner, 1976, 1984; Galloway, 1989), ii) episodic tectonism (Copeland, 1959; Bott and Johnson, 1967; Klein and Willard, 1988, among others), and iii) climate change (Cecil, 1990).

It is improbable that established paleomires were terminated by river avulsion; if so, why are the Springhill seams unfailingly overlain by several metres of sideritic mudrock? Rather, it seems that

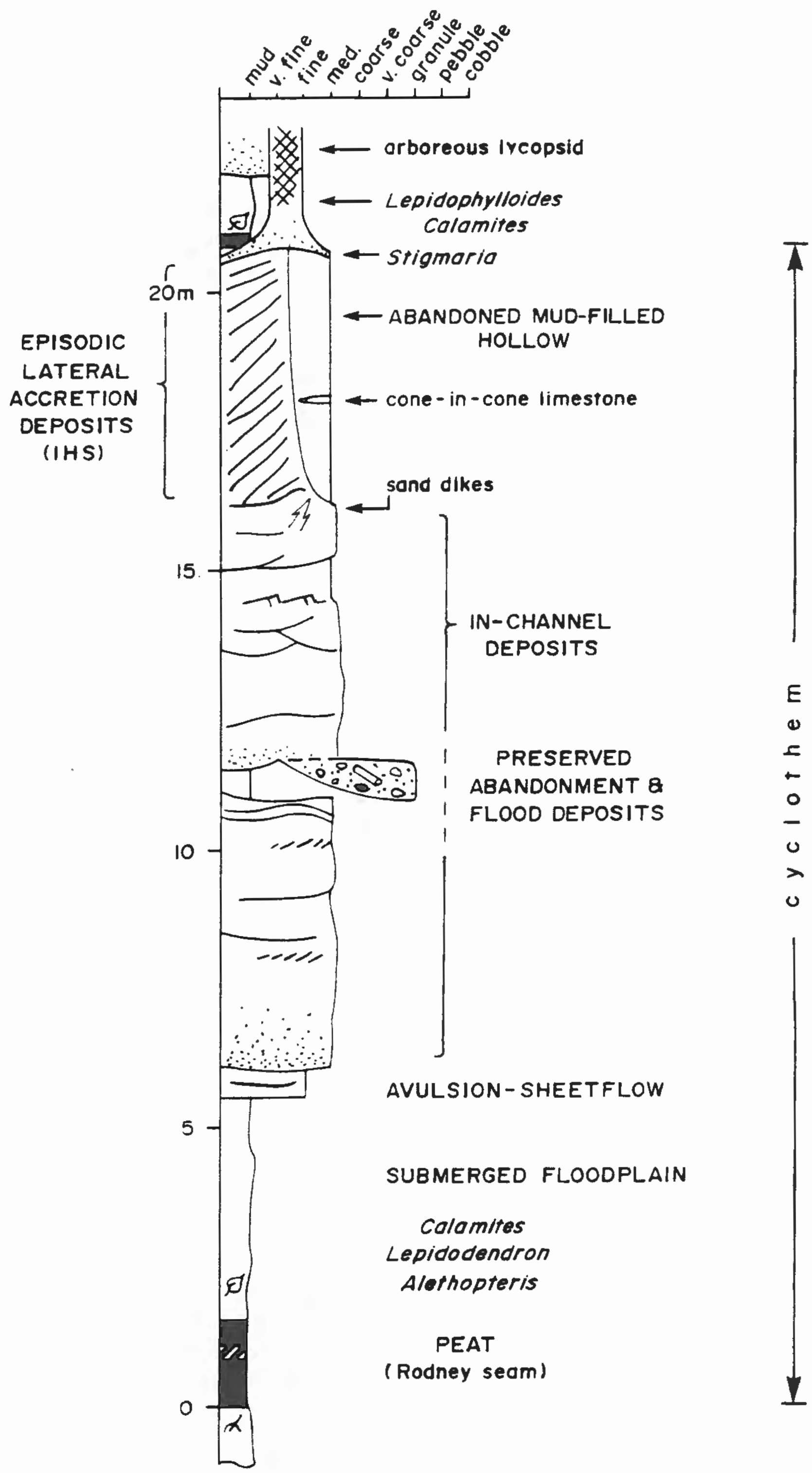


Figure 3.15 A cyclothem of the Springhill coalfield, comprising the Rodney seam and Rodney Sandstone.

the river migrated in *response* to mire burial. It is probable that mire life was influenced by conditions including peat accumulation and fertility (Clymo, 1987) in balance with basin subsidence rate (Fulton, 1987). In searching for a possible allogenic cause, the prerequisite of a regularity of process would appear to eliminate episodic tectonism. Climatic cyclicity, however, may have exerted an influence and will be investigated further in the concluding chapter.

3.8.4 Cycles of Intraseam-Scale:

Cycles of intraseam scale include sequential development of lithotypes and seam partings comprising siliciclastic sediments which interrupt lithotype development. Although the thickness of coal between intervening partings is commonly of the metre- to decimetre-scale, siliciclastic splits in the seam may be an order of magnitude thicker, as revealed by the northerly 22 m-thick parting within the No. 3 seam (Figure 3.4). The presence of a 13 m-thick multistorey sandstone body within this split along the basin-axis side of the seam is significant in that the vertical arrangement of lithofacies within the split defines a cyclothem, which would otherwise be considered to be of a higher order of cyclicity.

Lesser partings where a seam borders either the basin margin or axis record the imposition of depositional processes of neighbouring deposystems (alluvial fans and axial rivers), which may have been triggered by episodic ephemeral events such as unusually severe rainfall or possibly by basin adjustment.

3.9 THE COAL WINDOW: EVIDENCE FROM THE BASIN-FILL SEQUENCE OF A FUNDAMENTAL ALLOGENIC CONTROL ON MIRE DEVELOPMENT

Upon examination of the basin-fill sequence of the Springhill coalfield (Figure 3.14), it is readily apparent that coals are not randomly interspersed in vertical profile, but are preferentially associated with grey mudrocks and multistorey sandstones of assemblage V which occurs above the lacustrine-fill of assemblage III and below the reddened mudrocks and grey multistorey sandstones of assemblages V-VI. The preferential occurrence of coals beneath such redbeds was earlier noted by Hacquebard and Donaldson (1964). A window of peat (coal) formation during basin development is therefore evident, apparently constrained by a maximum and minimum water level.

In evaluating this hypothesis, a method is proposed wherein three attributes of the basin fill are considered: sandstone body thickness; occurrence of variegated (mottled) and predominantly red mudrocks; and coal seam thickness. By plotting these three parameters in vertical section (Figure 3.16), it is possible to interpret relative groundwater level, and subsequently to infer causes. Changes in groundwater level may be induced by several factors including: i) tectonism (gross subsidence rate), ii) the rate at which sediment infills the basin as it subsides (net subsidence rate), iii) climate, and iv) eustasy.

This method of basin analysis is dependent upon the following assumptions:

i) Sandstone body thickness and number of storeys varies in proportion to subsidence rate. This is due to an increased proportion of multistorey bodies as the channel-belt combs across the alluvial plain and amalgamates with underlying bodies. It is assumed that periodicity of avulsion remains constant within an order of magnitude (Allen, 1978).

ii) Mudrock colour reflects groundwater level. Reddish and variegated mudrocks form at or above the water table (Bown and Kraus, 1987), particularly where the groundwater level fluctuates (Robinson, 1949). Grey, sideritic mudrocks represent hydromorphic soils (Besly and Fielding, 1989) formed under perennially saturated conditions (Reading, 1980; Besly and Fielding, 1989), whereas reddish mudrocks represent relatively well drained soils (Besly and Fielding, 1989). Given constant climatic conditions, reddish mudrocks should reflect relatively emergent conditions and improved drainage (Besly, 1988) which prevail under conditions of relatively low net subsidence rates.

iii) Occurrence of coal, in particular that formed from rheotrophic, groundwater-influenced mires, is dependent upon groundwater level. Peat formation was impeded by unsuitably low or high groundwater levels induced by variations in subsidence rates. With prohibitively high levels of groundwater arising from excessively rapid subsidence, the peat-forming ecosystem will have drowned, whereas with very low groundwater levels, the accumulating plant matter will have decomposed under aerobic conditions. In extreme conditions where the ecosystem is dependent upon groundwater virtually as the sole source of recharge, lush vegetation may not be sustainable.

At present there is considerable disagreement in the literature as to the effects of subsidence on alluvial architecture. Some computer models and field occurrences suggest that channel belts migrate to areas of more rapid subsidence (Bridge and Leeder, 1979; Read and Dean, 1982; Leeder and Alexander, 1987; Alexander and Leeder, 1990) resulting in a higher density of sandstone bodies in such areas. Thickness of sandstone bodies may be affected, however, by factors such as river planform, and confinement either by narrow valleys or by mud-rich floodplains. Increased sandstone body thickness may be brought about by downcutting/incision of a channel belt, especially in the case of anastomosed rivers (e.g. Rust *et al.*, 1984). In such cases, the vertical spacing of bodies may increase also. The model, therefore, may require modification as controls on sandstone body density and thickness become better understood.

Kraus (1987) observed a direct relationship between pedogenic maturity and sandstone body thickness and multistorying which she inferred to indicate decrease in basin subsidence. Kraus (1987, p. 609) further observed that "In tectonically active basins,....pedogenic maturity should decrease during periods of tectonic activity or rapid subsidence and increase during periods of tectonic quiescence or slowed subsidence". The colouration of mudrocks as used herein does not address maturity of paleosols, rather it is a matter of different pedogenic processes and soil types. The implication is, however, that pedogenesis within the basin-fill is a potentially important indicator of groundwater level and by inference of net subsidence rate.

The occurrence of thick coals implies persistent optimum conditions for peat accumulation. Their presence is a phenomenon which must be explained in light of the inferences drawn from interpretation of groundwater level. In summary, this method considers direct evidence of groundwater level, from which subsequent inferences can be drawn with respect to causes such as net subsidence or eustasy.

3.9.1 Springhill Coalfield Basin-Fill Sequence

The thin, unistoried sandstone bodies of lacustrine affinity and poorly developed coals in the lower 180 m of the Springhill coal-bearing sequence below the Gesner seam reflect a relatively high water level (Figure 3.16). Lithofacies assemblage III, which comprises this interval, is interpreted

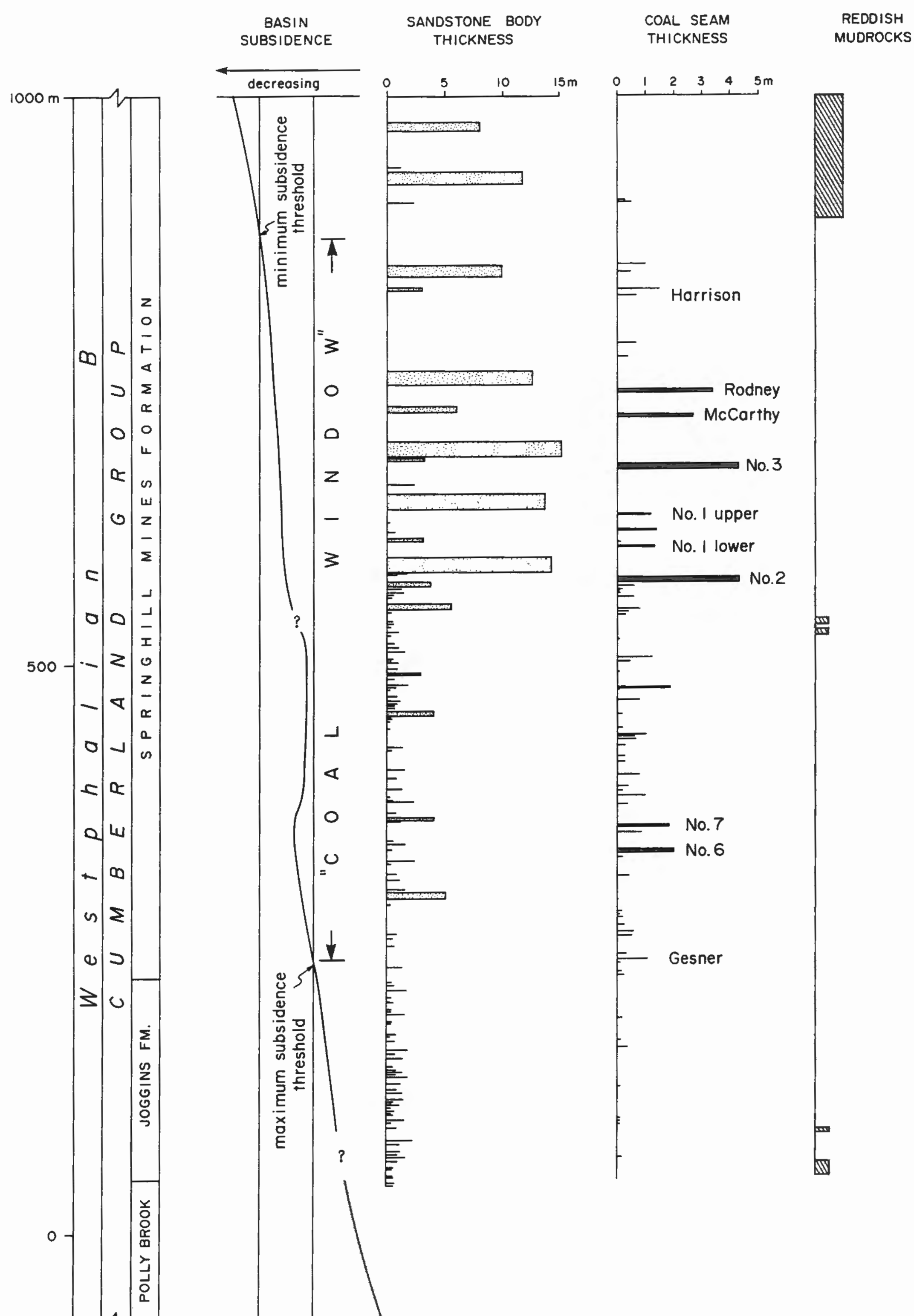


Figure 3.16 Vertical distribution of sandstone body thickness, coal seam thickness and red beds (mudrocks) in the Springhill coalfield basin-fill sequence illustrating the coal window, inferred to have been groundwater-controlled.

(Chapter II) as a lacustrine/bay-fill assemblage. The occurrence of strata of lacustrine affinity is consistent with the interpretation of a relatively high base level and high rate of basinal subsidence (Blair, 1987b; Blair and Biloudeau, 1988).

A somewhat lower groundwater level prevailed throughout the depositional history of the overlying 300 m of strata. Although the Gesner seam marks the approximate time when groundwater level dropped sufficiently for thick peat accumulation, groundwater level continued to be high, near maximum for the allowable "coal window". A possible exception was during deposition of the Nos. 6 and 7 seams and associated multistorey sandstones (Figure 3.16).

Optimum conditions for thick peat development coincided in part with those resulting in multistorying of sandstone bodies. This interval extends from the No. 2 seam to the Rodney seam, a stratigraphic distance of 200 m. Here, coal seam and multistorey sandstone body thickness reaches 4.3 m and 15.2 m respectively. Where an appropriate groundwater level coincided with a favourable net rate of subsidence for thick peat and erosive amalgamation of sandstone bodies, as in this interval, thick coals resulted. As infilling of the basin continued, however, the groundwater level became lower and peat accumulation declined, although the thickness of sandstone bodies was maintained or increased. This phase of basin evolution is recorded by the decline in coal seam thickness above the Harrison seam, and ultimate absence of coals once groundwater level fell below the minimum threshold: reddening of mudrocks occurred concurrently with this decline.

Although a progressive decline in groundwater level is apparent within the basin-fill, the underlying causes have yet to be established. Potential causative factors are changes in sediment supply, crustal loading and subsidence, salt removal, and orogenically induced climate change associated with denudation of the Cobequid Highland Massif.

3.10 THE COAL-BEARING BASIN-FILL SEQUENCE AT JOGGINS (LOWER COVE TO RAGGED REEF)

Although this study deals primarily with the coal-bearing strata of the Springhill coalfield, it is useful to contrast the basin-fill sequence exposed on the Joggins shore. Unfortunately, the Joggins section, albeit a classic exposure of Carboniferous strata (Gibling, 1987), has not been appraised in detail in modern times except on a piecemeal basis (Duff and Walton, 1973; Rust *et al*, 1984, Salas, 1986). Comparison of the two basin-fill sequences is therefore constrained, particularly with respect to interpretation of evolving depositional environments recorded by the basin-fill.

The basin-fill column of the Joggins section from Lower Cove to Ragged Reef (Figure 3.17) was compiled from the reference measured section of Logan (1845) as given by Dawson (1855). It incorporates all of Divisions 4 and 3 of Logan, and the lowermost part of Division 2. The section spans an areal distance of 6 km perpendicular to the dip of the strata. The 1430 m coal-bearing sequence is underlain by a 635 m thick section dominated by reddish mudrocks (Division 5 of Logan, 1845). The lower 800 m of the coal-bearing sequence (Division 4 of Logan, 1845) includes thin coals commonly overlain by limestones bearing a bivalve fauna, grey and red mudrocks and single storey and less commonly multistorey sandstone bodies. Together, these basin-fill entities comprise the Joggins Formation (Bell, 1914, revised by Ryan *et al.*, in press). The overlying 630 m of the coal-bearing sequence (Division 3 of Logan, 1845), named the MacCarrons Creek Member and assigned to the Springhill Mines Formation by Ryan *et al.* (1990; in press), differs from the underlying coal measures in the absence of bivalve-bearing limestones and in the greater abundance and thickness of multistorey sandstone bodies.

The systematic upward change in lithological attributes recorded in the coal-bearing portion of the basin-fill sequence at Springhill (Figures 3.14, 3.16) is less obviously developed within the Joggins section due to the superimposition of smaller-scale cycles, including red beds. Broad similarities nevertheless exist between the two sequences: strata of lacustrine affinity are common in the lower portions of both, and both exhibit an ultimate upward reddening of mudrocks, decrease in thickness of coal seams and increase in thickness and abundance of multistorey sandstone bodies. The Joggins coal-bearing sequence differs from that of Springhill, however, in several key aspects:

i) the coal seams seldom exceed 1 m in thickness at Joggins;

ii) the presence of ubiquitous bivalve-bearing limestones which occur in association with, and commonly overlying, the coal seams of the Joggins Formation at Joggins (Division 4 of Logan);

iii) there is less affinity between thick, multistorey sandstone bodies and relatively thick coals at Joggins than at Springhill;

iv) reddened mudrocks comprise a higher proportion of the coal-bearing sequence at Joggins as opposed to Springhill;

v) the thickest interval devoid of red mudrocks at Joggins is 180 m, between

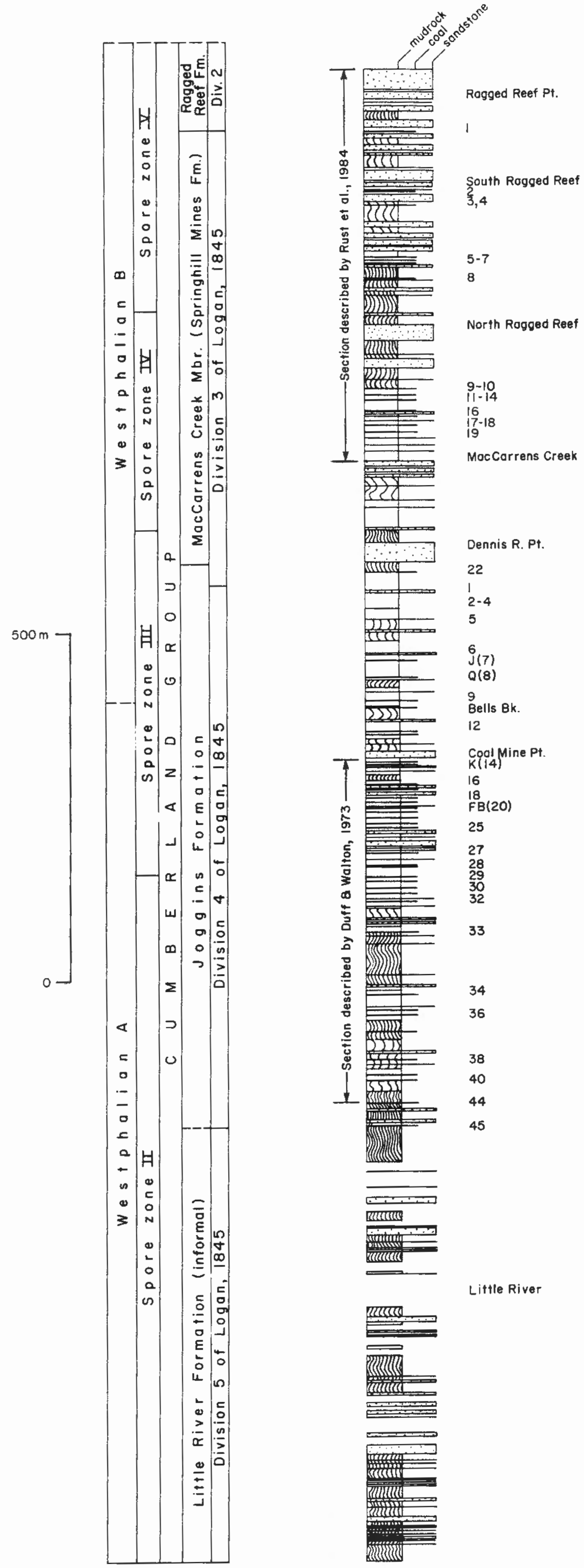


Figure 3.17 The Joggins section, drawn from the measured section of Logan (1845).

coals 33 and 16 of Division 4 of Logan in comparison with 430 m at Springhill, beneath the No. 2 seams (indeed if it were not for the minor occurrence of mottled mudrock beneath the No. 2 seam, fully 790 m of the Springhill sequence would be devoid of redbeds); and

vi) the coal-bearing sequences are underlain by extrabasinal conglomerates at Springhill and by a red mudrock-rich section (Logan's Division 5) at Joggins.

The upward trends in redbeds, coal and sandstone thickness were recognized within the upper 650 m of the Joggins sequence (MacCarrons Creek Member) by Rust *et al.* (1984). They attributed the upward increase in reddened mudrocks and decrease in coals to progressively lower groundwater levels, a conclusion supported by the author. They interpreted the upward change in sandstone thickness (and density) to represent a change from a meandering to an anastomosed channel pattern. They inferred that the evolution of the sequence was a response to rapid subsidence and to increased slope of the depositional surface. These trends could alternatively be explained by relatively higher *net* subsidence to the east (basinward) of the MacCarrons Creek part of the Joggins section, or by a eustatic drop in sea level resulting in downcutting and regionally lower groundwater levels, as postulated for coal-redbed cycles in the Sydney coalfield (Gibling and Bird, in press).

Discussion:

The basal deposits of the southern basin margin at Springhill differ markedly from those at Joggins. The occurrence of predominantly grey strata of lacustrine affinity at the base of the Springhill coal-bearing sequence would appear to suggest greater basin subsidence there than at Joggins, hence basin asymmetry. Seismic data acquired by the Geological Survey of Canada (Bromley and Calder, in press) and biostratographic correlation of Dolby (1991), however, indicate that the coal-bearing sequence thickens northwesterly from Springhill to Joggins, paradoxically suggesting greater subsidence at Joggins (see also Figure 3.12). A similar increase in stratigraphic thickness is evident in the coal-bearing strata of the Joggins-Chignecto coalfield from Chignecto west to Joggins (Copeland, 1959).

The apparently coeval deposition of red beds at Joggins with conglomerates of the basal Cumberland Group near Springhill may be attributed to a number of causes, including: i) a semi-arid climate during the Westphalian A, (Dolby, 1991); ii) the red beds occupied a position on the relatively uplifted side of the basin, with alluvial fans deposited on the down-dip side as hypothesized by Alexander and Leeder (1987). Subsequent deposition at Joggins coeval with later fan building would have taken place under somewhat higher water conditions, although of fluctuating level; or iii) alluvial fan and red bed deposition occur during periods of relatively low subsidence rate, as suggested by Blair (1987) and Blair and Biloudeau (1988).

The late stages of alluvial fan building at Springhill were marked by a retreat of the fans, indicated by the vertical change from assemblage I to II of predominantly braided fluvial to sheetflow deposition. Sheetflow deposition on alluvial fans occurs predominantly below the intersection point where channels cease to be entrenched within the fan apex (Denny, 1967; Hooke, 1972; Bull, 1972, 1977; Kesel and Lowe, 1987). The reverse vertical arrangement is indicative of fan progradation (Blair, 1987b).

As fan deposition entered its later stages at Springhill, the environment marginal to the fans evolved from shallow lacustrine to an alluvial plain with major channel belt and mire development (Nos. 6 & 7 seams). With retreat of the alluvial fans at Springhill, deposits of the alluvial plain onlapped the distal fans. Peat mires occurred preferentially adjacent to the basin margin and reached their developmental zenith both in terms of individual mire longevity and frequency of mire development. The basin-fill record at Joggins at the same time, however, reveals a decline in the longevity of mires and increased reddening of floodplain muds. As onlap of the alluvial plain deposits of the southern basin margin ensued and the basin filled, the area of submerged wetlands and floodplain became increasingly restricted, as groundwater supply diminished. Drainage eventually improved to a point where gleying of alluvial plain muds was widespread and peat virtually ceased to accumulate.

The basin-fill sequence manifested in the lithofacies assemblages of the Springhill coalfield records decreasing tectonic subsidence at a basin margin reflected by retreating alluvial fans and increasing contribution from distally-derived sediment carried by longitudinally flowing fluvial systems. Some of the through-flowing rivers may have had their source near a tectonically active constraining bend of the Cobequid fault to the southwest, although the timing of deformation there is poorly constrained (Plint and van de Poll, 1984). Samples obtained by the author from the Gardiner Creek and Tynemouth Creek Formations in southern New Brunswick yielded a Westphalian B palynomorph assemblage correlative with spore zone 3, the Joggins Formation (Dolby, 1988). Other contributions of longitudinally-derived sediment may have come from as far away as the central Appalachian collisional orogen (Gibling *et al.*, in press).

Both the Springhill and Joggins basin-fill sequences reflect a change to progressively drier environments of deposition. Differences in the sedimentology and interpreted environments of the various components reflect relative position and prevailing tectonic setting within the basin, and access to groundwater recharge, in particular from basin-margin alluvial fans.

The mutual occurrence at Joggins of bivalve-bearing limestones and thinly bedded mudrock and sandstone of lacustrine affinity together with reddened mudrocks which formed under emergent or near-emergent conditions is an apparent enigma. This apparently contradictory association may have arisen from fluctuating groundwater levels on a very low-gradient alluvial to shallow lacustrine region. Relatively minor vertical change in water level thereby resulted in areally widespread flooding or drying of the topography, resulting in the drowning of mires and formation of bivalve-bearing strata, or reddening of mudrocks, respectively. The allogenic control responsible for such variation in water level has yet to be identified, awaiting further study of the Joggins section, but possibilities include orbitally-forced climatic change (Milankovitch cycles), discussed in Chapter V, sediment supply, and perhaps eustasy. It is not known whether Joggins occupied a northern basin margin, but it is apparent that the area was more distal to alluvial fans than was the Springhill area. The Joggins area, therefore, was less likely to have benefited from supplemental groundwater discharge from fans during period of lowered groundwater level, whatever the underlying cause.

3.11 CONCLUSIONS: ALLOGENIC INFLUENCES ON WESTPHALIAN B PEAT ACCUMULATION IN THE CUMBERLAND BASIN

The allogenic influences of tectonism and geomorphology affected the development of the peat mires chiefly through distribution of groundwater. Climate effects, also influential, are discussed in greater depth in Chapter V. From the evidence secured to this point, several key points can be made concerning the mires and associated strata of the Springhill coalfield which determine the type of mires and certain prerequisites to their formation:

1) thickest coals occur preferentially with multistorey sandstone bodies, especially those for which the periodicity of channel belt revisitation (Allen, 1978) was relatively short;
2) major coals occur preferentially adjacent to distal fan toes;
3) no cases have been identified at Springhill of major coals occurring other than between the two deposystems above;
4) the topographic position, sedimentology and petrography of mires suggests that they received meteoric groundwater recharge from shallow subsurface distal fan discharge,

and from ephemeral surficial distal fan sheetflow and perhaps also from basin-axis meander belts during high stage flow;

5) thick peat accumulation occurred as basin-floor deposits onlapped retreating fans;

6) early mires (below No. 6 seam) and those distant from the southern basin margin (as at Joggins) were vulnerable to water-table fluctuation; late mires (above Rodney seam) saw decreasingly favourable sites for accumulation (i.e. perennially saturated), indicating an optimal window of mire development during basin infilling;

7) the window of mire development within the basin-fill sequence is inferred to represent a period of favourable net basin subsidence, possibly coincidental with (?orogenically induced) climatic change.

8) similar orientation of channel-belt and peat deposits may be coincidental but suggests faults may have been active, exerting some influence on local topography, subsidence, groundwater flow, hence river course and peatland area.

The preferential development of peatlands relatively close to a basin margin, specifically adjacent to alluvial fans, is in apparent contradiction with the theory that economic coals formed in sites far removed from active deposystems (McCabe, 1984). It is probable that the ancient mires of the Springhill coalfield were able to develop adjacent to the distal fans only after the alluvial fans had become established landforms permitting aquifer development and associated sheetflooding had reached a stage whereby the mire vegetation could baffle incoming sediment and groundwater. The relationship between alluvial fans and peatlands, and the coal window developed during basin filling, emphasizes the importance of groundwater supply in the development of certain mires, ancient and modern.

By definition, therefore, the ancient peat mires of Springhill therefore would have developed as *rheotrophic* (flow-fed) ecosystems (Moore, 1987) fed both by precipitation and groundwater. The modern bog forests of Sarawak and Brunei, often cited as analogues of Carboniferous mires (e.g. McCabe; 1984, Moore, 1987; Grady *et al.*, 1989; Grady and Eble, 1990) are in contrast ombrotrophic, fed by precipitation (Anderson, 1964; Moore, 1987). These late Carboniferous basin-margin coals of the Springhill coalfield are significant in that they appear to have developed as rheotrophic mires whereas Moore (1987) stated that low ash, wood-derived Carboniferous coals "can only have developed in an ecosystem of a tropical "forest" type in which the abundant supply of water was entirely by precipitation, i.e. ombrogenous."

CHAPTER IV: MIRE GENESIS: THE NO. 3 SEAM

4.1 INTRODUCTION

The analogy of raised peatlands (mires) as precursors of coal seams is the center of a debate that is fundamental to interpretations of the origin of coal. Apart from being one of the most interesting problems to be solved in the future (Teichmüller, 1989), it is of fundamental importance not only from an academic standpoint but also in the development of exploration strategies, which draw heavily upon interpreted controls on the deposition of ancient peat, hence coal. In addition, the nature of the developing, precursor peatlands influences the type and distribution of sulphur and other mineral matter within a seam, the paleoflora that contributed to the accumulating biomass, and the genesis of derived coal macerals.

The debate, in essence, is whether raised, (domed) mires, and in particular tropical bog forests, are widely applicable as an analogue for ancient coal seams. The concept was introduced in an important study by Smith (1962) who drew an analogy between modern raised peatlands of tropical Malaysia and Indonesia and the Westphalian coal seams of Yorkshire, U.K. The debate has developed as geologists investigating coal and coal-bearing strata readily embraced the concept of the raised mire and applied it to the ancient, at times in absence of objective, definitive criteria. Such conclusions were supported by the views of peat researchers, such as Moore (1987): "such [low ash, wood-derived] material can only have developed in an ecosystem of a tropical 'forest' type in which the abundant supply of water was entirely by precipitation, i.e. ombrogenous." And, while clearly such tropical bog forests are suitable precursors of low ash coals, there is troubling contradictory evidence from many coals (Teichmüller, 1989), one of the most obvious problems being the intercalations of flood-derived siliciclastic partings within many coal seams (Teichmüller, 1982). As observed by Moore (1989, p. 90), "there has been an unfortunate lack of information exchange between coal geologists and ecologists concerned with modern peatlands, to the detriment of both areas of study."

This chapter focuses on the genesis of ancient peat-forming ecosystems that were precursors to the coals of the Springhill coalfield. In particular, this portion of the research addresses the apparent contradiction between evidence secured from sedimentologic, stratigraphic and basin analysis, which suggest a fundamental groundwater control on mire development, and the prevailing theory that thick, low-ash coals formed from raised (domed), rain-fed mires (McCabe, 1984; Moore, 1987, among others). This chapter considers the validity of methods employed in assessing ancient peat-forming systems, particularly from the standpoint of raised (ombrotrophic) versus rheotrophic (cf. planar) mires. It is fundamental that the tools for the analysis of ancient peat deposits (coal) be selected or developed from those employed in the classification of modern peatlands. The three avenues of approach in the interpretation of ancient peat-forming systems that will be addressed in this chapter are lithotype, maceral and miospore analysis. The case study utilized herein is the No. 3 seam of the Springhill coalfield.

4.2 CLASSIFICATION OF PEAT-FORMING ECOSYSTEMS

In this study, the term *mire* is employed for any freshwater wetland system in which peat accumulates (Moore, 1989). The nomenclature and main definitive criteria employed in the classification of the mire types *bog, fen* and *swamp* (Gore, 1983; Moore, 1987) are summarized in Table 4.1. No inference is drawn in this study to plant composition that typifies modern examples of bog, fen and swamp.

The principal attributes by which mires are characterized are: 1) groundwater influence (Kulczynski, 1949; Moore and Bellamy, 1974; Grosse-Brauckmann, 1979; Gore, 1983; Etherington, 1983); 2) nutrient/ionic supply (Kulczynski, 1949; Moore, 1987; Gore, 1983); 3) ash content (Gore, 1983; Moore, 1987); and 4) vegetation type (Gore, 1983; Moore, 1987). The generalized developmental trends of major mire types with respect to these attributes is depicted in Figure 4.1. Hydrologic character, including groundwater influence and source of ionic input, is both a pragmatic and ecologically meaningful basis on which to classify mires (Moore, 1987). Similarly, the two primary variables in the development of peat-forming plant communities are water and ionic supply (Kulczynski, 1949 and Tallis, 1983), which are intrinsically related. Changes either self-induced (autogenic) or externally induced (allogenic) in these environmental variables result chiefly from modifications to hydrology or surface topography of the mire (Tallis, 1983). A fundamental distinction of mires, therefore, is between **ombrotrophic** mires, which are solely rain-fed, and **rheotrophic** mires, which receive recharge from both groundwater and rainfall. Rheotrophic mires are usually minerotrophic and eutrophic (nutrient-rich), whereas ombrotrophic mires are usually oligotrophic (nutrient-poor). Ombrotrophic, rain-fed mires are invariably raised and domed (ombrogenous); the commonly used geomorphic terms **domed** and **planar** are loosely analogous to the hydrologic terms ombrotrophic and rheotrophic, respectively (see Cecil *et al.,* 1985). A term useful in interpreting the temporal evolution of a mire, whether modern or ancient, is **mesotrophic,** which refers to a mire tending toward ombrotrophic conditions. A fundamentally important point for geologists concerned with the interpretation of ancient mire development is that many domed

CRITERION:	W E T L A N D E C O S Y S T E M S			
Peat Accumulation:	PEAT - FORMING **(MIRES)**			NON PEAT-FORMING
Water Source:	RAIN-FED (OMBROTROPHIC)	FLOW-FED (RHEOTROPHIC)		
		Dry Season Water Table		
		Below Surface	Above Surface	Near Surface
Water Level:	**Bog**	**Fen** [1]	**Swamp** [2]	**Marsh** [3]
Vegetational varieties:	(bog forest)		(floating swamp, swamp forest)	

[1] sensu stricto as orginally proposed by Tansley (1939). Moore and Bellamy (1973) and Gore (1983) applied fen in the broader sense to all rheotrophic mires.

[2] "swamp" in the United States commonly refers to forested wetlands (Penfound, 1952) without regard to accumulation of peat. Those swamp forests which are peat-forming are similar to Scandinavian "carr".

[3] As orginally proposed by Tansley (1911). In United States, marsh and swamp are commonly differentiated by presence (swamp) or absence (marsh) of trees (Penfound, 1952).

Table 4.1 Wetland ecosystems and hydrological types of mires, based upon Moore (1987).

ombrotrophic mires *evolve* from a rheotrophic, planar state. The tropical, raised forest mires of Indonesia, for example, originated as groundwater-influenced mangrove soils (Polak, 1975; Anderson, 1964; Cameron *et al.*, 1989).

Rheotrophic and planar mires generally exhibit a higher ash content than do domed, ombrotrophic mires (Moore, 1987, his Figure 6). Mires can be subdivided into type on the basis of vegetative cover, so that, for example, a swamp forest can be recognized within the swamp type (Moore, *ibid.*). The reader is referred to Gore (1983) and Moore (1987, 1989) for elaboration of the processes and classification of peat-forming ecosystems.

4.3 THE RECOGNITION OF RHEOTROPHIC/PLANAR VS. OMBROTROPHIC/DOMED PALEOMIRES

The fundamental templates for the recognition of ancient peatlands types are those that characterize modern peat-forming systems (Figure 4.1). A mire tending toward a raised (ombrotrophic) status (cf. doming) should be marked by the following upward trends: 1) decreasing groundwater influence, hence 2) decreasing nutrient supply and ionic input, which can be expressed as decreasing fertility (Anderson, 1983); 3) decreasing content of water-lain ash; and 4) a tendency toward more acidic conditions (decreasing pH). The search for supportive or contrary evidence of these trends within the descendant coal seam is discussed at some length in this chapter.

Calder *et al.* (1991) discussed criteria for the recognition of rheotrophic and ombrotrophic conditions in coal seams, especially as they applied to maceral-based methods of analysis. They concluded that the long-standing use of inertinite as a relative measure of Eh level may be intrinsically flawed in determination of some paleomire types. The major reasons given were: 1) Eh level is little used as a definitive criterion of modern mire types; and 2) fusinite, a common inertinite maceral, *when of the variety pyrofusinite* owes its origin to wildfire, which is an event related to short-term climatic rather than geomorphic conditions (Cope and Chaloner, 1985; Collinson and Scott, 1987, Teichmüller, 1989; Scott, 1989). In addition, such fire-derived fusinite is particularly prone to transport by wind or consequent run-off (Scott, 1989), which further lessens its usefulness as a sensitive paleomire indicator. Moreover, as pointed out by Esterle *et al.* (1989), recent studies of modern tropical forest mires (Esterle *et al., ibid*; Grady *et al.*, 1989) have not revealed the upward increasing trend of inertinite precursors cited by numerous workers (Esterle and Ferm, 1986; Littke, 1987; Warwick and Flores, 1987; Eble and Grady, in press) as being indicative of a seam derived from

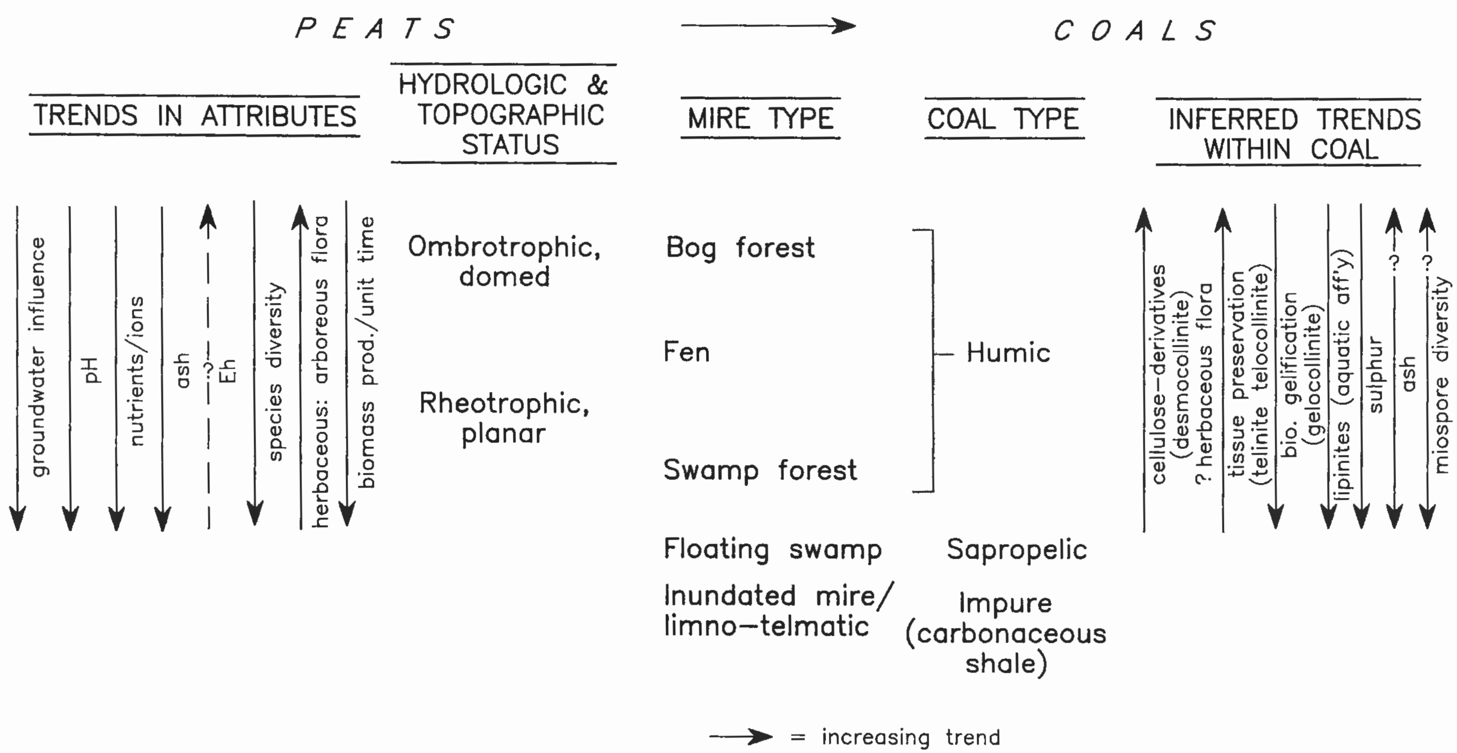


Figure 4.1 Generalized schematic trends of attributes of mire development for principal rheotrophic to ombrotrophic peat/coal-forming mire types.

a paleomire that evolved to a raised (domed) status. Some inert macerals may develop during diagenesis, which may account for their abundance in coal relative to peat and lignite (Teichmüller, 1989). Further studies into the genesis of inertinite of degradational origin will prove valuable in paleoenvironmental interpretation.

One criterion directly transferable from the modern to the ancient is the temporal (upward) decrease in (waterlain) ash content in mesotrophic mires. This trend may be exaggerated by the concomitant leaching of inorganic matter (Cohen, 1974, 1989; Cecil *et al.*, 1982; Cohen *et al.*, 1987) as the pH level of the mire decreases with raising of the surface relative to groundwater level. An increase in mineral matter content relative to organic matter has been inferred to accompany the transformation of peat to bituminous coal (Cecil *et al.*, 1982) but the vertical trend should remain. Although the introduction of flood-derived detritus into the mire is the result of an event (flood) as is wildfire, a mire that receives such input is by definition rheotrophic (groundwater influenced). The occurrence of a parting within a seam does not mean that the seam developed under the influence of groundwater throughout its entire history, however. Caution must be exercised as well in the interpretation of mires that were sites of volcanic ashfall, resulting in the formation of kaolinitic tonstein. (Tonsteins have yet to be identified from the coal-bearing strata of the Cumberland Basin.)

Nutrient and ionic supply and pH level can be deduced indirectly by considering their effects upon vegetation, maceral and mineral formation. Decreasing fertility of the tropical bog forests of Indonesia and Malaysia results in the development of areally zoned plant communities (Anderson, 1983), characterized by: 1) floral change (only one species of tree, *Shorea albida,* spans all six communities); 2) a reduction in species diversity; 3) an increase in stems per unit area; and 4) a decrease in average size of a species (stunting).

The first two of the changes described by Anderson can be measured directly in the ancient record through palynomorph analysis. The effects of an increase in the number of stems and stunting of trees (Anderson, 1983) on the development of coal are less straightforward. Stunting of vegetation in response to raising of the mire surface has been invoked to explain dulling-upward sequences in coals (Smith, 1962; Esterle and Ferm, 1986; Warwick and Flores, 1985). In a test of this hypothesis, however, Esterle *et al.* (1989) found that the modern tropical bog forests of Southeast Asia would give rise to a brightening upward lithotype sequence. An effect that has not been considered is the increase in number of stems or roots, hence outer bark, per unit area that accompanies stunting (Anderson, 1983). There are fundamental botanical problems with the analogy between Carboniferous coals and Indonesian peats, however. Carboniferous peats were composed largely of lycopsid bark; they contained almost no wood. Indonesian peats on the other hand, are virtually all wood. Furthermore, lycopsid bark was probably quite unlike that of angiosperms, especially in its chemistry and sheet-like character (DiMichele, written communication, 1990).

Changes in ionic supply and pH level within the developing mire have a profound effect not only on growing vegetation but also upon macerals and mineral matter. The development of vitrinite macerals in particular is strongly affected by these two variables (Teichmüller, 1982, 1989). A modern tropical domed peat from the Baram River region of Sarawak (Esterle *et al.*, 1989) is characterized by a macroscopic succession from sapric through hemic to fibric types (ASTM, 1987) with a decrease upward and peripherally in maximum size of woody fragments. Microscopically, plant tissues are better preserved upward, whereas the poorly structured 'matrix' decreases. Cellulose-derived matter (red textinite, attrinite, densinite) has been found to increase upward within the domed mires of Malaysia and Indonesia (Esterle *et al.*, 1989; Grady *et al.*, 1989), a reflection of floral change upward and inward to herbaceous vegetation (Anderson, 1964; Cameron *et al.*; 1989; Esterle *et al.*, 1989) in response to infertility arising from lessened groundwater influence and nutrient supply. Biochemical gelification of lignin-derived humins, promoted by the presence of water (particularly when oxygenated) and alkaline ions (especially Ca), results in the development of unstructured forms of vitrinite, in particular gelocollinite (Teichmüller, 1982, 1989). A mesotrophic paleomire tending from planar to domed would therefore be expected to reveal a vertical (temporal) decrease in the effects of biochemical gelification evident in the distribution of vitrinite maceral types, especially telinite, weakly structured vs. unstructured telocollinite and gelocollinite (Calder *et al.*, 1991). Such a trend may be largely independent of the kind of tissue entering the system, but confirmation of this awaits further research of degradation processes. An increase in cellulose-derived macerals, specifically attrital desmocollinite, may be expected, however, if trends toward stunted, herbaceous floras occurred as in modern mires (Figure 4.1).

4.4 METHODOLOGY: THE NO. 3 SEAM

The No. 3 seam was chosen for study because of its broad areal extent and considering the distribution and quality of data and the availability of samples in a north-south transect across its breadth (Figures 4.2, in rear pocket; 4.3). Samples were obtained from diamond drillcores chosen from the inner zone and north (riverine) and south (piedmont) marginal zones of the seam. Preference was given to those cores with the highest degree of recovery. Study involved:

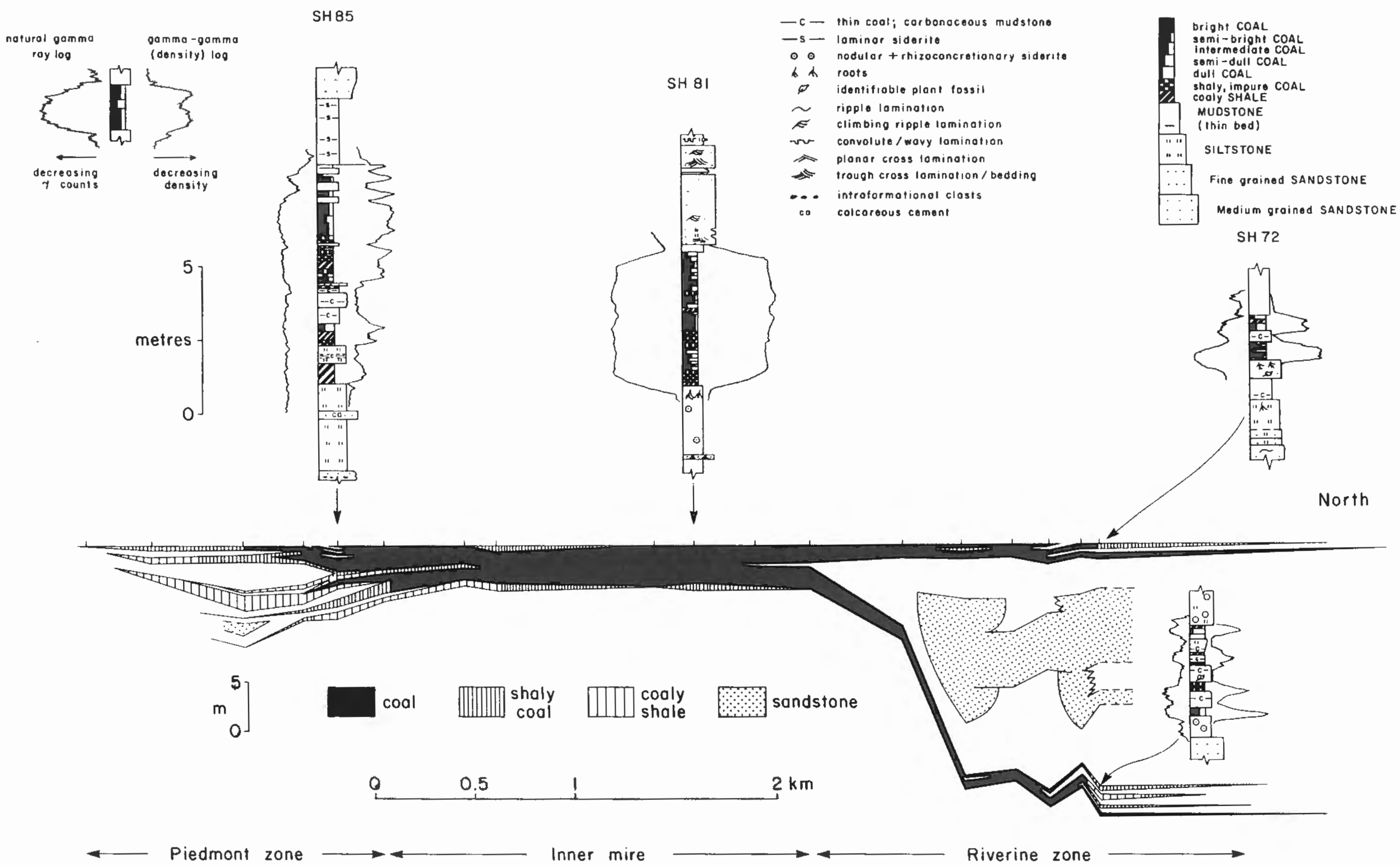


Figure 4.3 Basin margin to basin axis (N-S) cross-section of the No. 3 seam. Profiles are of drillcore used in lithotype, maceral and miospore analysis.

i) detailed macroscopic description and measurement (minimum individual lithotype thickness: 5mm);

ii) determination of location and volume of core lost in drilling, involving matching graphic log of seam petrography to 1:25 scale gamma and density signatures (e.g Figure 2.11);

iii) sampling of individual lithotypes or intervals of similar lithotypes (to avoid bias in areas of broken core, entire intervals, rather than splits, were taken).

Samples were submitted to the Atlantic Coal Institute, Sydney, Nova Scotia for stage crushing to 20 mesh. Each sample was split, with half of all samples submitted to Dr. Graham Dolby, Calgary for palynological (miospore) analysis. The remaining splits were used to manufacture particulate grain mount pellets at the Atlantic Coal Institute for maceral analysis.

Maceral analyses (volume percent) were conducted on 52 samples of the No. 3 seam by Dr. P. K. Mukhopadhyay, Atlantic Coal Institute. The samples were examined both under normal reflected (white) light and under blue-light excitation. Maceral analyses of the Rodney seam were earlier conducted under normal reflected light by the writer. The procedures employed are in accordance with those recommended by the International Committee for Coal Petrology (1963; 1971; 1975). Proximate and sulphur analyses (weight percent) were also conducted at either CANMET Labs or Atlantic Coal Institute on all samples examined for maceral content.

Nine composite samples representing the three areal zones of the No. 3 seam were submitted for trace element and major oxide analyses of ash (see Appendix). All 13 samples from drillhole SH 81 in the medial region of the seam were submitted for ultimate analysis of carbon, hydrogen and nitrogen (see Appendix) to X-Ray Assay Laboratories, Don Mills, Ontario and Neutron Activation Services, McMaster University, through the Atlantic Coal Institute. These data are presented in a separate publication (Goodarzi and Calder, in prep.).

4.5 GEOLOGIC ASPECTS OF THE NUMBER 3 SEAM

The geometry and extent of the Number 3 seam, (see Chapter III), are better understood than for any other seam of the Cumberland Basin. The seam (Figures 4.2, 4.3) attains a maximum thickness in its medial region of 4.32m (SH 81) and extends 6.5 km from northern to southern margin (transverse span). Its east-west (longitudinal) extent is unknown due to post-depositional erosion of the coal-bearing strata to the east and to the great depth of these strata to the west beyond the extent of drill and mine data; the minimum longitudinal extent is 4 km but is presumed to have been much greater as there are few signs of deterioration at either known extremity.

During deposition of the major coal-bearing lithofacies assemblage (V), as described in the preceding chapter, peat-forming ecosystems 4 to 7 km in width flourished along a margin of retreating coalesced alluvial fans derived from the Cobequid highland massif to the south. The northern margin of the peatlands was bordered toward the basin axis by the medial reaches of a river belt 1-2 km wide (Calder, 1986). The interaction of the mires and neighbouring geomorphic systems resulted in a distinct areal zonation of the thickest and most areally extensive seams, readily apparent in north-south cross-section (Figure 3.4). This zonation is best exemplified by the No. 3 seam. Three zones are apparent, defined by the lithology and geometry of the seams and siliciclastic splits (Figures 4.2, 4.3): 1) a ***piedmont zone*** to the south, bordering the basin margin; 2) an ***inner zone***; and 3) a ***riverine zone*** to the north, bordering the basin axis. The inner zone of the No. 3 seam is largely devoid of siliciclastic partings. The southern, or piedmont, zone is characterized by an abrupt lateral change in seam geometry and lithology. Siliciclastic partings (ill-sorted mudrocks, graded pebbly sandstone), rare in the inner zone, become numerous over a distance of 300 - 500m. The incorporation of these strata, which are individually of dm-scale but amalgamate locally to form partings greater than 2m thick, results in a marked increase in seam thickness. The riverine zone is marked both by dramatic seam splits around entombed multistorey sandstone bodies and by mudrock partings.

Hacquebard *et al.* (1967) recognized the preferentially mineable inner zone and the rapid lithologic change, which they termed lithification, that typifies the piedmont zone, although these areal terms were not used. Similar areal zonation and basin-fill sequences characterize intermontane coal basins such as those of the Mesozoic (Fuxin Basin) of China (Li Sitian *et al.*, (1984), and the Tertiary (Powder River Basin) of the U.S.A. (Obernyer, 1978), which are drained by through-flowing basin axis fluvial systems (Flores, 1989).

4.6 NO. 3 SEAM: LITHOTYPE ANALYSIS

More than 30 macroscopic seam profiles were recorded for the No. 3 seam, but those profiles chosen for comprehensive (lithotype, maceral and palynomorph) analysis are the focus of this section (whole core from drillholes SH85, piedmont zone; SH81, inner zone; SH72, riverine zone).

	PIEDMONT ZONE	INNER MIRE	RIVERINE ZONE
	(South, basin margin)		(North, basin axis)
GROSS GEOMETRY	. abrupt lateral change; . dm-scale, commonly ill-sorted mudrock and graded, pebbly sandstone partings becoming numerous over distance of 300-500m, giving rise to marked increase in seam thickness.	. relatively free of seam splits	. major splits (up to 22m thick) about entombed multi-storey sandstone bodies, with further splitting of major leaves;
TREND OF LITHOTYPES	. upward brightening of lithotypes.	. overall upward-brightening of lithotypes, with slight reversal in trend at top.	. variable trends in lithotype brightness.

		PIEDMONT ZONE (drillhole SH85)	INNER MIRE (drillhole SH81)	RIVERINE ZONE (drillhole SH72)
LITHOTYPE PARTICIPATION	. vitrain	3.0	1.6	0.9
	. vitroclarain[1]	1.5	4.6	-
	. clarain	7.2	42.4	16.3
	. duroclarain[2]/ clarodurain	33.8	31.3	21.5
	. durain	3.2	6.9	24.1
	. fusain	0.5[3]	5.3	0.4
	. impure coal	50.8	7.9	36.9

[1] syn. bright clarain
[2] syn. dull clarain
[3] locally as high as 13.5% (drillhole SH27)

Table 4.2 Summary macroscopic characteristics of areal zones, No. 3 seam, Springhill coalfield.

4.6.1 Areal Variation in Lithotype Abundance

The macroscopic petrography of the No. 3 seam reveals marked lateral (areal) variation in lithotype composition (Table 4.2). The middle, or inner, zone of the No. 3 seam is characterized by abundant clarain (42%) and duroclarain (31%) and a paucity of impure coal lithotypes (shaly coal, coaly shale: 7.9%). Within the riverine zone to the north, clarain decreases to 16% whereas durain and impure coal increase to 24 and 37% respectively. In comparison, the piedmont zone shows the greatest proportion of impure coal (51%). Although there is little difference in duroclarain between the piedmont (34%) and inner (31%) zones, clarain decreases sharply within the former (7%). Vitrain, as macroscopically recorded bands, comprises <3% of the seam in any zone. Within the three measured sections, fusain is most abundant in the inner zone (5.3%) but is locally highly abundant in the piedmont zone, comprising as much as 13.5% of the No. 3 seam in drillhole SH27 (Figure 4.2). The high percentage of fusain in this intersection can be attributed in part to a solitary 6 cm thick bed composed of erratically oriented cubes of fusain that bear a remarkable resemblance to charcoal.

4.6.2 Vertical Sequences (Trends) of Lithotypes

The macroscopic columns (Figures 4.3 and 4.5 to 4.7) exhibit an overall upward change in lithotypes. In the piedmont zone (SH85) the basal portion of the seam is composed predominantly of impure lithotypes, whereas the upper part of the seam is dominated by brighter lithotypes (chiefly clarodurain and duroclarain), which increase in abundance upward and thus confer an overall brightening-upward trend. A similar vertical trend is apparent in the inner zone, but with a slight reversal toward duller lithotypes in the upper third of the seam. The inner zone exhibits a greater proportion of brighter lithotypes throughout and virtual absence of siliciclastic partings. The riverine zone to the north shows variable trends with no clear, overall motif to the vertical arrangement of lithotypes.

4.6.3 Preferred Sequence of Lithotypes: Markov Chain Analysis

In order to investigate whether lithotypes exhibit a preferred vertical sequence, the macroscopic data from the three zones were subjected to Markov chain analysis (Harbaugh and Bonham-Carter, 1970). The preferred pathways of upward transition from one lithotype to another are presented for each zone in Figure 4.4. In all three cases, two major populations of lithotypes are apparent based on their preference for preceding lithotypes.

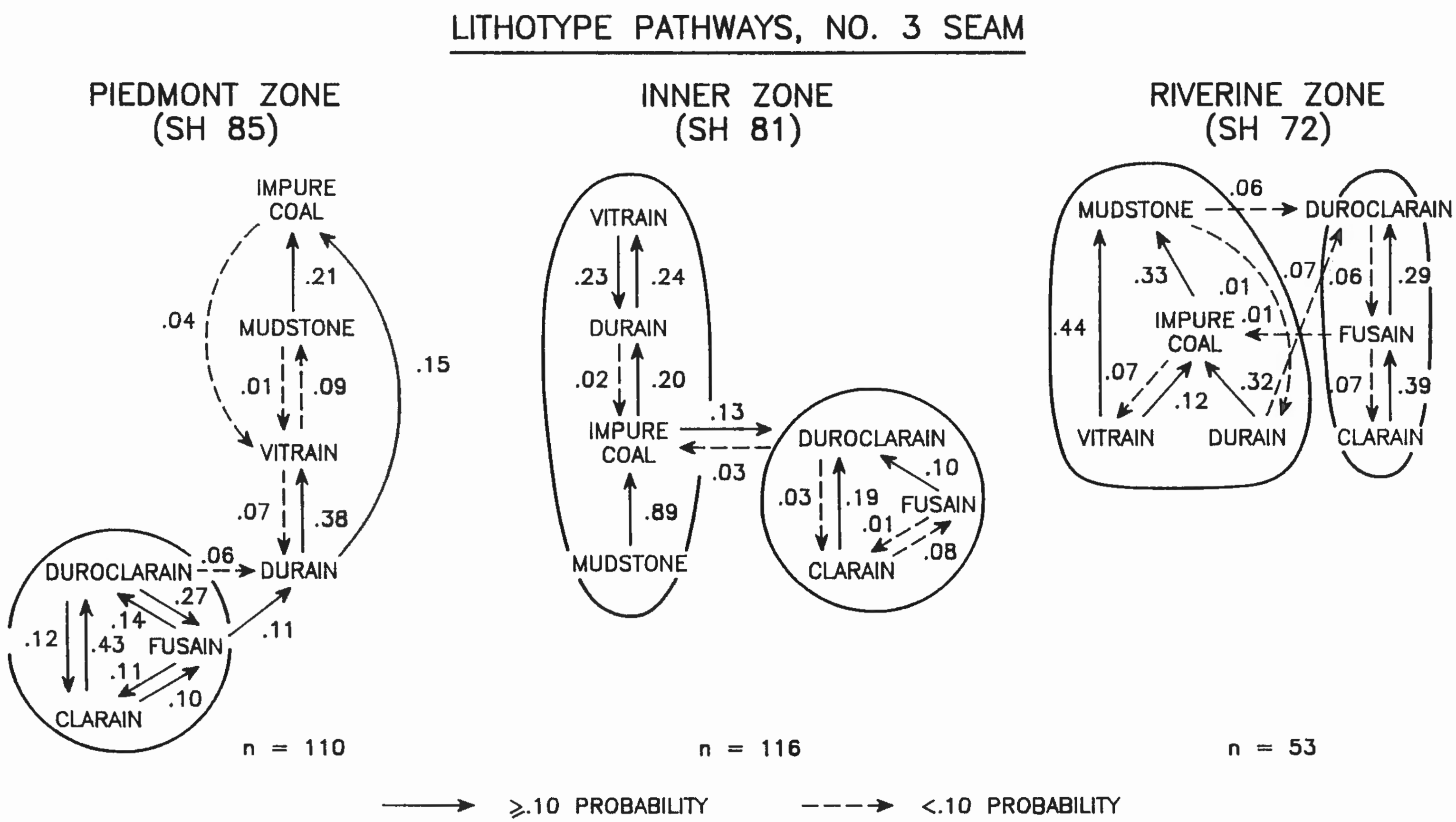


Figure 4.4 Preferred vertical succession pathways of lithotypes from the three areal zones of the No. 3 seam, as deduced from Markov chain analyses.

These two groups are 1) clarodurain/duroclarain/clarain/fusain; and 2) mudrock/impure coal/durain/vitrain. Upward transitions within the second of these lithotype groups are rather varied from zone to zone (Figure 4.4). Hence, this group describes an inter-relationship among this lithotype population rather than a vertical succession. The inner zone shows less variation or stronger preference in pathways than either marginal zone and the least number of cross-transitions between lithotypes of the two groups.

4.7 INTERPRETATION OF LITHOTYPE DISTRIBUTION

Two lithotype populations, rather than one linear sequence (Tasch, 1960), were identified through Markov chain analysis. These reflect two broad environments of peat accumulation: the "dry" population (fusain, clarain, duroclarain, clarodurain) represents the established mire, whereas the "wet" population (mudrock, impure coal, durain, vitrain) reflects the heavily flooded, inundated mire. Fusain and clarain are closely associated as preceding states in the Markov chain analysis. Fusain, representing fossil charcoal, is believed to be the record of wildfire (Cope and Chaloner, 1985; Scott, 1989), which is essentially an "event". Such fire horizons may be more valid as an indicator of short-term climatic change than of topography within the mire. Vitrain, the brightest lithotype, occurs within the "wet" population of lithotypes in the No. 3 seam. This seeming paradox is attributed to the incorporation in mud of thick, discrete vitrain layers or bands representing bark sheets of arboreous lycopsids, which favoured wet habitats.

According to the summary of lithotype abundance by zone (Table 4-2) and the interpretation of lithotypes (Table 4-3), both mire margins developed under generally higher groundwater levels than did the inner zone. The riverine zone to the north witnessed predominantly wetter (limnotelmatic) conditions with the deposition of durain and impure coal. Minor pathways indicated in the Markov chain analyses between "wet" and "dry" communities (Figure 4.4) suggest fluctuating groundwater level, but at a century scale given the average lithotype thickness (4.8 cm) and assuming peat accumulation ratios of 2.3 mm yr^{-1} as in modern tropical forest mires (Anderson and Muller, 1975) and a coal:peat compaction ratio of 7 to 1 (averaged by Ryer and Langer, 1980). Although this zone was proximal to the major river channel belt, which formed under strongly fluctuating discharge, the temporal scale of flood deposits preserved within the fluvial and mire deposits may have differed. Systematic upward changes on a small (laminar) scale were poorly developed, largely due to the interruption of autogenic peat-forming processes by random, allogenic events and perhaps due to the interference of autogenic peat evolution by differential compaction around entombed channel-belt sandstone bodies. The stronger pathways of the inner mire suggest that peat accumulation there was

TASCH (1960)		CALDER (THIS STUDY)					
LITHOTYPES	RELATIVE GROUNDWATER LEVEL	LITHOTYPES	PLANT COMMUNITY: TER	S–TER	TEL	L–TEL	MIRE CONDITION
FUSAIN	DRY	FUSAIN	—	—	- - -		(WILDFIRE)
BRIGHT COAL AND BRIGHT BANDED COAL (VITRAIN AND VITROCLARAIN) BANDED COAL (CLARAIN)	↑	CLARAIN	—	—			ESTABLISHED MIRE
DULL BANDED COAL AND DULL COAL (CLARODURAIN AND DURAIN)		DUROCLARAIN CLARODURAIN		—	—		(TRANSITIONAL)
CARBONACEOUS SHALE	↓ WET	DISCRETE VITRAIN[1] DURAIN IMPURE COAL		- - -	- - -	—	FLOODED, INUNDATED MIRE

TER: TERRESTRIC S–TER: SEMITERRESTRIC TEL: TELMATIC L–TEL: LIMNOTELMATIC

[1] within mudrock

Table 4.3 Lithotype interpretations of Tasch (1960) and Calder (this study).

steadier, and less interrupted by episodic events. The piedmont zone to the south shows a strongly bimodal composition of duroclarain/clarodurain and impure coal. The general rarity of cross-over between the wet and dry lithotype populations of the piedmont zone (Figure 4.4) may indicate that these represent temporally distinct habitats.

In order to obtain a true picture of seam (mire) evolution, it is necessary to inspect the entire seam column for the overall upward sequence of lithotypes. The overall upward-brightening tendency of lithotypes within seam columns of the inner mire and piedmont zones (Figures 4.2, 4.3 and 4.5-4.6) indicate that there was a progressive evolution of the mire throughout much of its history and over much of the peatland area. It can be deduced from the vertical succession of lithotypes (Figures 4.5-4.7) that the seam originated as a flood-prone mire. As the ecosystem evolved, successive organic accumulates record a relative drop in groundwater level upward and toward the piedmont as the mire became established and accreted vertically and laterally. This suggests that the mire was mesotrophic, i.e. tending toward less groundwater influenced conditions. A reversal toward duller lithotypes in the uppermost part of the inner mire column (SH81), however, suggests a rise in groundwater level relative to the mire surface. The northern margin of the mire was throughout its existence the site of episodic riverine inundation, which interfered with long-term evolution of the mire.

In summary, four main inferences can be drawn from the lithotype analysis of the No. 3 seam: 1) two broad environments are reflected by the lithotype population: i) established mire, and ii) flooded, inundated mire; 2) systematic upward change was interrupted by allogenic random events; 3) steadier, uninterrupted peat-forming processes are indicated by the stronger pathways of the inner mire; and 4) the overall trend of lithotypes evident in the macroscopic seam columns indicates a progressive, mesotrophic development of the mire throughout its history, particularly in the inner mire and piedmont zone.

4.8 MACERAL-BASED PALEOMIRE ANALYSIS

Factors that must be taken into account in the maceral-based paleoenvironmental interpretation of a coal seam include: 1) the botanical precursors of the maceral; 2) transport of macerals within the paleomire; 3) diagenetic history, including physiochemical change; and 4) secondary macerals whose precursors are unknown. The trophic status of the mire profoundly affects not only the contributing flora but the preservation and transformation of the accumulating peat. This is particularly significant within the vitrinite maceral group, which is especially susceptible to diagenetic transformation due to the chemistry of the contributing plant matter.

4.8.1 Botanical Precursors

The clearest botanical affinity exists for the exinitic macerals of the liptinite group, because they have undergone less diagenetic transformation than certain other macerals. These exinitic macerals are derived from plant exines and epiderm. They include spores and pollen (sporinite), cuticles of leaves and areal plant extremities (cutinite) and algal bodies (alginite). Suberinite is derived from suberin in corkified bark tissues.

Liptodetrinite is derived from finely detrital particles of other liptinite macerals; it may have multiple botanical precursors. Resinite is derived primarily from resinous secretions in tree barks (terpene resinite) such as those of *Cordaites sp.* and *Sigillaria sp.* (Teichmüller, 1982). Such resinite is usually associated with telinite (Plate 2.18). The potentially important ecological affinity of terpene resinite is diminished however by the less diagnostic occurrence of secondary resinite (cf. fluorinite of Mukhopadhyay *et al.*, 1979) and lipid resinite derived from waxes and fats.

Vitrinite and inertinite group macerals commonly share similar botanical origins (see Table 4.4); the resulting maceral is dependent in large part on Eh level. The bulk of vitrinite is derived from lignin and cellulose, principally the former (Teichmüller, 1982; Stout and Spackman 1987). Inertinite macerals for the most part share similar parent materials with the vitrinite macerals, although certain macerals such as sclerotinite are unique to this group. Sclerotinite may form either from fungal mycelia which flourished in high stress environments, or from oxidized resins (Kosanke and Harrison, 1957; Cook and Taylor, 1962; Stach, 1966). Sclerotinite of either origin may be indicative of oxic conditions, but the fungal type could also result from waterlogged or toxic environments (Teichmüller, 1982).

4.8.2 Diagenetic Transformation of Vitrinite Macerals

Of the three maceral groups, vitrinite macerals are most susceptible to diagenetic change due to their chemistry. Since macerals of the vitrinite group commonly constitute the bulk of Carboniferous coal seams, diagenetic transformation of the plant precursors poses a significant hindrance to interpretation of the affinity of the macerals within the context of mire ecology.

Plant Precursor	Site of Origin	Process/Pathway	Maceral, Type
lignin-rich secondary xylem of woody plants (especially barks of Lycophyta and woods of Gymnosperms) (also from thin-walled "reeds")	forested mires	humification, gelification	**telocollinite, tellinite**
	(+also in "reed" peat)	fusinitization (oxidation by charring, fungal degradation, etc.)	**fusinite, (pyrofusinite,+ degradofusinite), semifusinite**
gel precipitate from colloidal humic substances	cavities; peat-oxic water interface	gelification	**gelocollinite**
		oxidation	**macrinite**
cell excretions secondary cell fillings as humic gel	woody peats	humification gelification	**corpocollinite**
degraded humic detritus chiefly cellulose; low lignin	herbaceous mires	humification, degradation, maceration	**desmocollinite vitrodetrinite**
fungal mycelia, (cell secretions: see above - resinite and secretion sclerotinite)	high stress environments (subject to drought, waterlogging toxic conditions); 'reed' peats	primary fusinitization	**sclerotinite**
spores, pollen (sporopollenin)	all regions	n/a	**sporinite**
cutin of leaves stems, aerial plant parts	all regions especially telmatic to terrestric zones	n/a	**cutinite**
planktonic algae	open ponds	n/a	**alginite**
suberin-impregnated aerial woody parts ("cork")	forested mires	n/a	**suberinite**
resins of xylem ducts; lipid waxes	forested mires	n/a	**resinite**
		fusinitization (weathering or charring)	**secretion sclerotinite**
exines or lipid components	all regions	maceration	**liptodetrinite**

Table 4.4 Origin of macerals (based in part on Teichmüller, 1982 and Diessel, 1986)

The major processes of transformation are humification and gelification. In humification, the plant lignin is transformed in the peat acrotelm (upper, active layer) to humic acid substances, chiefly through microbial activity. Through diagenesis, these substances subsequently lose their acid character and are transformed into humins.

In gelification, the changes are largely physical in nature (Teichmuller, 1982), with the humins first brought into colloidal suspension (i.e. peptidization), and subsequently precipitated as a gel and desiccated. Peptidization is promoted by the presence of water, oxygen and alkaline conditions, especially Na and Ca ions (Teichmüller, ibid). Virtually all vitrinite macerals have been subjected to varying degrees of early diagenetic (biochemical) gelification. Differing degrees of biochemical gelification, depending on mire chemistry and original plant matter, result in a gradational series of vitrinite macerals from those that retain well-defined tissue structure (telinite) through a partial retention of structure (telocollinite) to those which have been completely gelified to a mobile, amorphous substance (gelocollinite, corpocollinite). Yet another vitrinite maceral (desmocollinite) arises from partially gelified, hydrogen-rich cell attritus that is lignin-poor (Teichmüller, 1982; 1989). (In very late stages of diagenesis, all huminites of the lignite stage are transformed to vitrinites of the bituminous coalification stage through geochemical gelification, otherwise known as vitrinization).

Vitrinite group macerals therefore record not only the type of parent plant matter - lignin-rich (telocollinite, telinite) versus cellulose-rich (desmocollinite) - but also the degree of early diagenesis experienced by the material, and related to water level, oxygen supply and relative pH levels (Teichmüller, 1989).

4.8.3 Transport of Macerals

An important factor that must be addressed in a maceral-based paleoenvironmental study is possible transport of macerals from their original site of formation (*in situ* versus allochthonous). The interpreted transport of macerals is perhaps the most subjective of the four factors. A fundamental premise is that macerals that form in a topographically low area of a mire with high groundwater level are unlikely to be transported to a higher region of the mire, whereas macerals formed at higher elevations have potential to be transported to lower areas (Diessel, 1982). Alginite and the "detrital macerals" (liptodetrinite, inertodetrinite, vitrodetrinite) are thought to accumulate in high groundwater (limnetic) areas (Diessel, 1982; Teichmüller, 1982; Mukhopadhyay, 1989). The author has observed, however, that algal mats occur in association with ponds in raised areas of recent domed mires in

Nova Scotia. Sites of accumulation for the macerals sporinite and cutinite of the liptinite maceral group and fusinite/semifusinite of the inertinite maceral group are contentious. Sporinite and cutinite have been used as indicators of a terrestric environment (Mukhopadhyay, 1986), however, the preferential site of accumulation for the originating spores is thought to have been the limnotelmatic zone (Diessel, 1982). The association of sporinite, alginite and mineral matter in the No. 3 seam maceral analyses supports the latter view.

The prevailing interpretation of fusinite and semifusinite has been that these oxidized plant remains formed in a relatively dry, terrestric paleoenvironment and that they are essentially *in situ* (Diessel, 1982; Harvey and Dillon, 1985; Mukhopadhyay, 1986). Fusinite and semifusinite can, however, accumulate in areas of relatively high groundwater level (Mastalerz and Smyth, 1988). Thus the occurrence of fusinite and semifusinite more accurately reflects the *input* of oxidized matter rather than an oxic environment at a given locality. In this study, the occurrence of fusinite and semifusinite in samples high in ash and alginite content supports the observation of Mastalerz and Smyth (*ibid.*).

4.8.4 Secondary macerals

Secondary macerals, i.e. derivatives of original macerals subsequently produced during coalification, do not reflect original environmental conditions within the mire. Such macerals should therefore be excluded during paleoenvironmental interpretation (Diessel, 1982). Secondary macerals identified in this study (Table 4.5) are exsudatinite, amorphous liptinite, and micrinite.

4.9 MACERAL ANALYSIS OF THE NO. 3 SEAM

The procedures employed in the maceral analyses are in accordance with those recommended by the International Committee for Coal Petrology (1963, 1971, 1975). Particulate grain mounts in lucite were manufactured, following stage grinding to -20 mesh, at the Atlantic Coal Institute, Sydney.

The point-to-point distance(s) between counts is determined according to the equation

$$S = 1/2\ d_{max} \text{ (Mackowsky, 1982)}$$

which for a maxiumum particle diameter of 1 mm obtained from a-18 mesh sieve (ISO coal preparation standard) is 0.5mm. The maximum particle diameter from a -20 mesh seive is 0.85mm. The maceral which occurs below the intersection of the single cross-hairs was then recorded. A point

MACERAL GROUP	MACERAL/MACERAL TYPE
Vitrinite	telocollinite/telinite
	gelocollinite
	corpocollinite
	desmocollinite
	pseudovitrinite "A"
	pseudovitrinite "B"
Inertinite	fusinite
	semifusinite
	macrinite
	sclerotinite
	inertodetrinite
	micrinite[2]
Liptinite	sporinite
	cutinite
	suberinite
	resinite/fluorinite[2]
	liptodetrinite
	alginite
	amorphous liptinite[2]
	exsudatinite[2]

[2] denotes secondary maceral

Table 4.5 Macerals recorded in the petrographic analysis of the No. 3 seam, Springhill Coalfield.

count of 500 is considered to be statistically valid using this method (Mackowsky, 1982). Such a count should be reproducible within limits of ± 1.5% with the same pellet surface and observer, according to Mackowsky (*ibid.*). Patterson and Fishbein (1989) have shown, however, that the smaller the fractional abundance of a constituent, the higher the number of counts required to maintain statistically acceptable variation. Maceral point counts were converted to percentage, which for maceral analysis of particulate grain mounts can be considered volume percent.

Maceral rather than microlithotype analysis was chosen as the preferred method of investigation for two reasons: i) due to areas of broken core, polished blocks, hence a constant microlithotype column, were not possible; ii) detailed subdivision of the macerals, in particular those of the vitrinite group, holds strong potential for in-depth evaluation of peat conditions; and iii) recent methods of paleoenvironmental assessment utilizing macerals (Diessel, 1982, 1986) were to be tested and if necessary modified as a new technique.

The macerals determined are listed by group in Table 4.5. Two macerals of particular note are suberinite and the pseudovitrinite type "B". The recognition of suberinite within the No. 3 seam constitutes the first report (Calder *et al.*, 1991) from a Carboniferous coal; previously, the maceral was known only from Tertiary and a few Mesozoic coals (Teichmüller, 1982) although it reportedly (C.F.K. Diessel, written communication, 1989) occurs in Permian coals of Australia. The identification of pseudovitrinite derived not only from telocollinite but also from gelocollinite has implications for the diagenesis of vitrinite. These findings are jointly reported with Dr. P.K. Mukhopadhyay, Global Geoenergy Research Ltd..

Maceral and mineral matter contents (volume percent) from analysis of the three seam sections are given in the Appendix. Since mineral matter is a constituent of coal, it is important to report the analyses on a "whole rock" basis. In the marginal zones of the seam however, large-scale trends within the maceral groups (see below) may be masked by dramatic variation in mineral matter content, hence these areas are described both in terms of a total coal basis (t.c.) and a mineral-free basis (m.f.).

As described previously, entire intervals were sampled from the drill core. Inorganic lithofacies which constituted 1 dm-thick partings were sampled (see for example the seam profile for SH72, Figure 4.7) and were subjected to palynological but not maceral analysis.

	Maceral Group Participation			Maceral Participation		
Zone	Trend	Range(%)	Proportion	Most Abundant	Range (%)	Remarks
Piedmont (SH85)	V increases upward	(16.0-74.2)	$V \alpha \frac{1}{\text{M.M.}}$	V: 1. gelocollinite 2. telocollinite 3. corpocollinite	(10.0-54.2) (3.8-18.6) (0-24.2)	- gelocollinite > telocollinite throughout - greatest at top
	L relatively consistent, decreases upward ultimately	(2.0-21.8)	L α I except:	L: 1. sporinite 2. alginite 3. resinite	(1.0-14.8) (0.6-4.4) (0.8--3.0)	- greatest at base - decreases upward
	I increase upward	(3.4-39.4)		I: 1. fusinite 2. semifusinite 3. inertodetrinite	(1.0-24.2) (0.2-5.0) (0-5.6)	- trend independent of fusinite
	M.M. decreases upward	(3.8-77.0)				- greatest at base
Inner (SH81)	V increases upward but declines at midseam	(39.2-79.4)	$V \alpha \frac{1}{I}$	V: gelocollinite in lower half, telocollinite in upper half	(5.8-37.6) (10.2-42.4)	gelocollinite > telocollinite to mid-seam, telocollinite > gelocollinite above.
	L decreases to mid-seam then increases; greatest at base	(7.2-22.6)	$L \alpha \frac{1}{\text{M.M.}}$	L: 1. sporinite 2. alginite 3. cutinite	(2.4-9.8) (1.0-6.0) (0.2-3.0)	decrease to mid-seam, then increase above greatest at base
	I variable, declines mid-seam	(8.0-27.6) except: L ≥ I at base	I > L	I. 1. fusinite 2. inertodetrinite 3. semifusinite	(0.6-10.8) (2.2-9.2) (2.8-8.4)	decline mid-seam
	M.M. declines mid-seam, greatest at base					
Riverine (SH70, 72)	V greater in upper leaf	(8.2-81.4)	$V \alpha \frac{1}{\text{M.M.}}$	V: 1. gelocollinite 2. telocollinite 3. corpocollinite	(3.4-42.8) (3.0-26.0) (1.2-15.0)	gelocollinite > telocollinite corpocollinite locally > telocollinite
	L decreases, then increases in both leaves; greatest at base of both	(7.8-25.8)		L: 1. sporinite 2. cutinite 3. alginite	(4.8-9.2) (0.4-6.2) (0.4-3.2)	- greatest at base - rare in lower leaf - increases upward, both leaves
	I 2-3 times greater in upper leaf	(1.4-54.0)	I α M.M.	I: 1. fusinite 2. inertodetrinite 3. semifusinite	(0.2-31.0) (0-14.0) (0.2-6.2)	greater in upper leaf
	M.M. lower in upper leaf but high at top	(2.2-65.0)				- increases upward in lower leaf

Table 4.6 Summary of maceral analysis from paleogeographic zones of the Number 3 seam (V = vitrinite, L = liptinite, I = inertinite, M.M. = mineral matter)

The seam sections show great variation in maceral composition. These variations occur as a hierarchy of trends, with smaller scale, lower order trends superimposed upon those of larger scale, higher order. A summary of trends and relative abundance of maceral groups and macerals is given in Table 4.6.

4.9.1 Maceral Abundance Profiles: No. 3 Seam

The maceral analysis of the No. 3 seam has been described by Calder *et al.* (1991); lithotype and maceral abundance of the three seam sections are depicted in Figures 4.5, 4.6 and 4.7. Calder *et al.* found, in summary, 1) predominance of the vitrinite maceral group in all three areal zones; 2) unstructured vitrinite (unstructured telocollinite and especially gelocollinite) predominated over structured forms (telinite and weakly structured telocollinite) throughout, except for the upper half of the seam within the inner mire; 3) abundant liptinite (in particular sporinite and locally alginite) at the seam base; 4) distinct vertical zonation of macerals (increasing telinite, telocollinite; decreasing gelocollinite and liptinite) and ash content (decreasing upward with slight reversal near top of seam) within the inner mire; and 5) petrographically distinct leaves (Nova Scotian coal mining vernacular: cf. 'benches' of U.S. mining terminology) in the riverine zone, including a paucity of structured telinite, telocollinite and inertinite within the lower leaf in comparison with the upper leaf.

4.9.2 The Record of Groundwater Influence in Peat and Coal Beds

The effects of two of the four templates for modern peatland classification, namely groundwater influence and nutrient/ionic supply, are reflected in the distribution of types of the vitrinite maceral types, which record varying degrees of biochemical gelification and contribution of lignin-rich vs. cellulose-rich tissues (Teichmüller, 1989). Biochemical gelification, influenced strongly by mire water chemistry, and the resistance of tissues (cellulose vs. lignin) to both physical and biochemical degradation are two main factors in the degradation of peat.

The degree of biochemical gelification of lignin-derived humins is largely a function of groundwater influence, for it is promoted by the presence of water, O_2 and alkaline ions. Biochemical gelification and loss of tissue structure is enhanced by bacterial activity, fostered by the increase in pH level of mire waters arising from the incorporation of alkaline ions (Teichmüller, 1982). Strongly gelified, unstructured vitrinite (especially gelocollinite) derived from lignin would therefore be expected to decrease upward in a rising mire as groundwater influence diminishes. Calder *et al.*

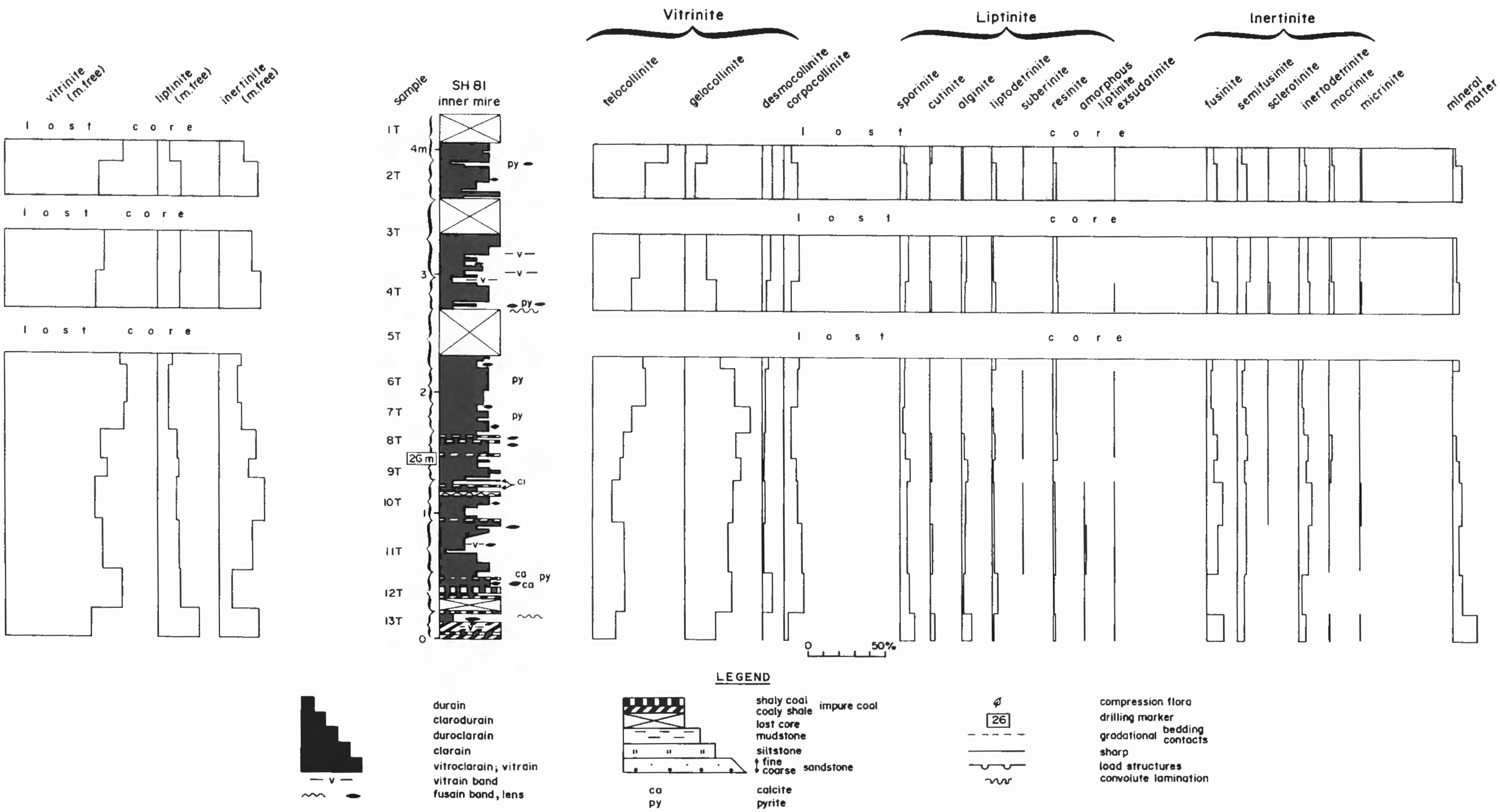


Figure 4.5 Maceral abundance profile of the inner mire zone, No. 3 seam (drillhole SH81).

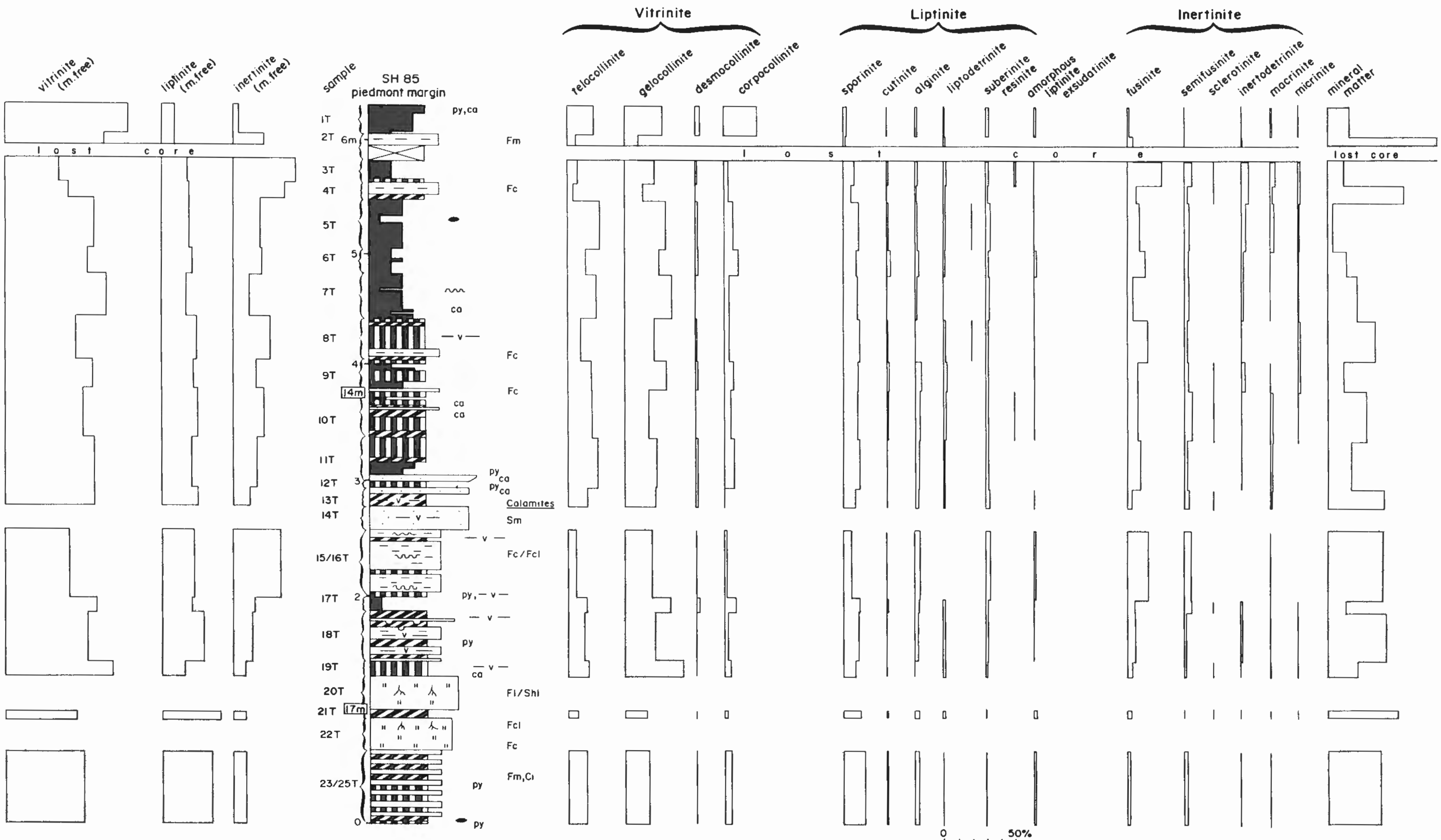


Figure 4.6 Maceral abundance profiles of the piedmont zone, No. 3 seam (drillhole SH85).

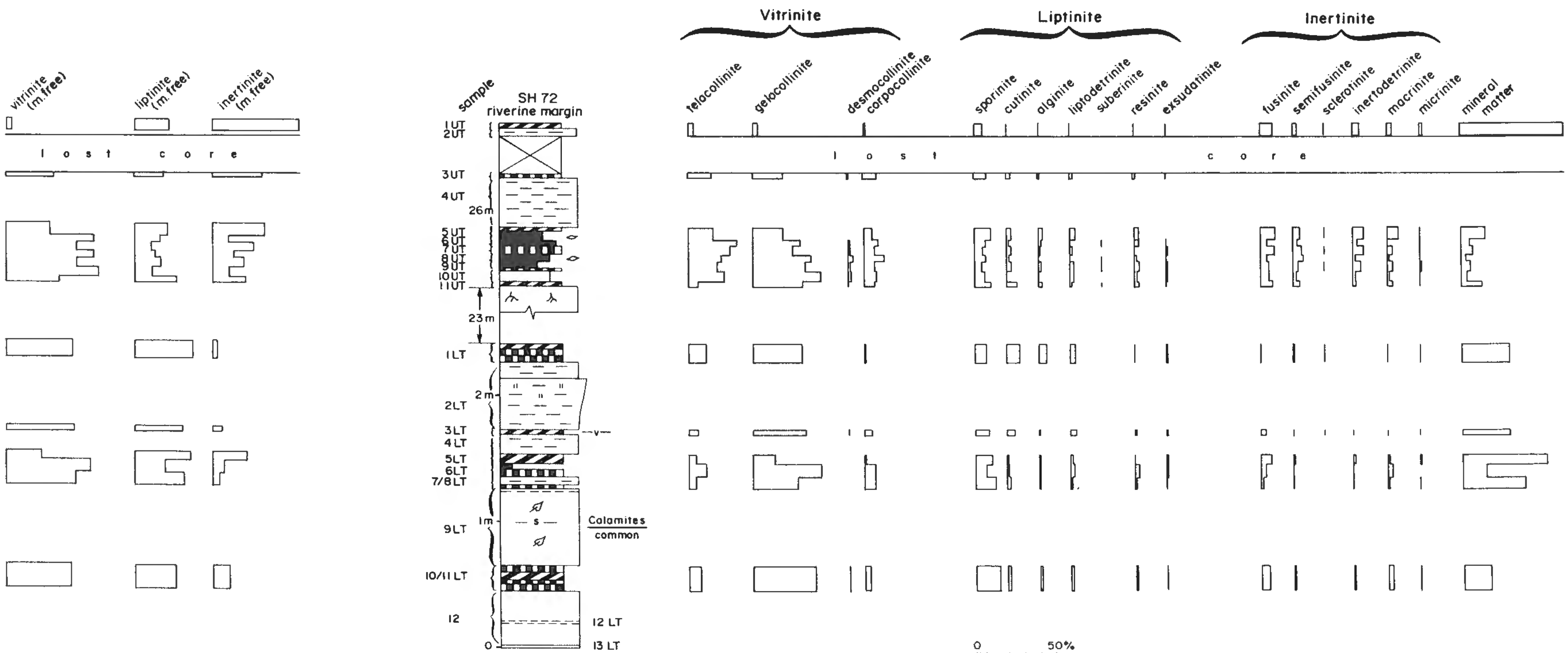


Figure 4.7 Maceral abundance profile of the riverine zone, No. 3 seam (drillhole SH72).

(1991) therefore stressed the need in studies of mire genesis to identify the collinite maceral types, and in particular to differentiate maceral types that reflect the gradual loss of tissue structure that accompanies progressive gelification (Koch, 1960; Timofeev and Bogoliubova, 1964).

Cellulose-rich tissues, presumably derived at least in part from herbaceous vegetation, are more prone to physical degradation (fragmentation) than are woody, lignin-rich tissues. The attritus forms humodetrinite and desmocollinite in coal (Styan and Bustin, 1983; Teichmüller, 1989). The occurrence of cellulose-rich, herbaceous vegetation in modern tropical mires is linked to decreasing fertility and nutrient supply that accompanies rising of the mire surface and diminishing groundwater influence (Anderson, 1983). Cellulose-derived macerals, precursors of desmocollinite (Teichmüller, 1989), have been found to increase upward within domed tropical peats of Indonesia and Malaysia (Esterle *et al.*, 1989; Grady *et al.*, 1989).

The physical and chemical pathways of degradation may conflict: assuming a similar tendency to herbaceous vegetation within ancient ecosystems, desmocollinite would increase upward within a raising mire, whereas strongly gelified vitrinite (e.g. gelocollinite) theoretically would be expected to decrease upward. It is fundamental, therefore, to specify which aspect of degradation is being considered in order to more clearly comprehend the genesis of ancient mires.

For this reason, it may be advisable to consider desmocollinite separately from other maceral types in a measure of groundwater influence. Furthermore, modern herbaceous mires may not serve as close analogues to Carboniferous ecosystems. Truly herbaceous plants were rare in Carboniferous ecosystems (W.A. DiMichele, written commmunication, 1990). It is necessary to resolve the paleoecology of plants less resistant to degradation than, say, the presumably lignin-rich arboreous lycopsids and which may or may not have occupied an ecological niche similar to the herbaceous flora of modern tropical mires.

Petrographic analysis of etched blocks from the Tertiary Anderson and Smith coals, Powder River Basin (Moore *et al.*, 1990), confirms these two degradation pathways. Moore *et al.* described two genetic maceral groups derived by degradation: a) structureless degradation products of woody material (their Genetic Group IIA); and b) particulate huminitic components (their Genetic Group IIB). As hypothesized here, these commonly show opposing trends (Moore *et al.*, 1990, their Figures 8 & 9).

The validity of gelocollinite, which may possibly form at various stages of peat diagenesis (Teichmüller, 1982), as an indicator of groundwater influence is contingent upon its early formation. Evidence of very early gelification within the No. 3 seam is illustrated, however, in Plate 4.1a, b. Restricted oxidation halos at contacts with gelocollinite indicate that this matter was already in a compact gelified state at the time that the adjacent structured tissues (semifusinite) were partially oxidized, perhaps through the process of smouldering groundwater (note presence of pyrofusinite, Plate 4.1a). The *abundance* of gelified matter is not an absolute indicator of rheotrophic conditions. Direct comparison between seams must take into account regional differences in groundwater chemistry and mire flora. The latter becomes especially important if the seams differ sufficiently in age for the evolution of plants to have become a factor.

Ash content, a third template of modern peatland classification, is another measure of groundwater influence, assuming that it is largely waterlain. Mineral matter can form also through diagenetic and biogenic processes (Cecil *et al.*, 1979; Cohen, 1990) and can be introduced by means such as volcanic ashfall (Dewison, 1989; Ruppert *et al.*, 1989). Such modes of origin do not seem to be widely considered in modern peat classifications. It should be possible, however, to more accurately assess the degree of groundwater influence by the differentiation of waterlain ash from that of other origins.

4.9.2.1 Groundwater Index

The relative intensity of rheotrophic conditions therefore can be evaluated using the following ratio, referred to as the **groundwater influence index (GWI)**:

$$\frac{\text{gelocollinite} + \text{corpocollinite} + \text{mineral matter}}{\text{telinite} + \text{telocollinite} + \text{desmocollinite}}$$

The ratio of strongly gelified (gelocollinite, desmocollinite, corpocollinite) to weakly gelified tissues (telinite, telocollinite) should provide a relative measure of gelification and by inference, of water supply and pH. The name preferred for such a ratio is Gelification Index, but this term has been previously employed by Diessel (1986) in a ratio contrasting vitrinite with inertinite macerals. As described above, however, the inferred trends for the relatively strongly gelified tissues in desmocollinite and gelocollinite may oppose each other because they reflect, among other things, differences in orginal plant matter and pathways of degradation. The change to herbaceous vegetation

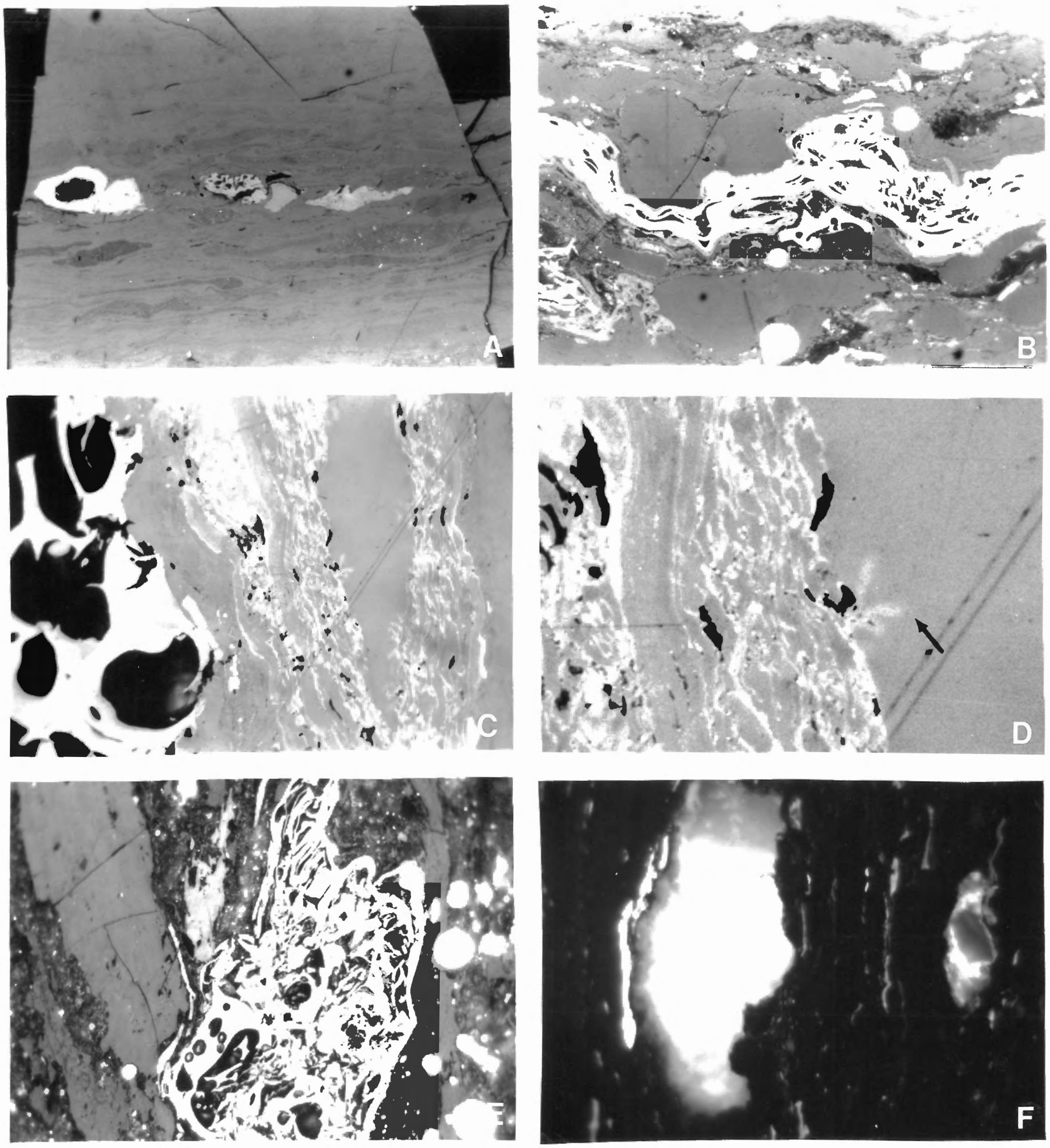


Plate 4.1 Photomicrographs from the Rodney and No. 3 seams in reflected light (x500 magnification: a) (?Fungal) sclerotinite bodies within telinite with resinite-filled cell lumens, Rodney seam; b) gelocollinite and degradofusinite, Rodney seam; c) pyrofusinite in contact with gelocollinite and semifusinite; semifusinitized tissues may represent smouldering through tissue tracts, No. 3 seam; d) enlargement of oxidation halo (arrow) at semifusinite/gelocollinite boundary; e) association of fusinite with degraded vitrinite, mineral matter and pyrite, No. 3 seam; f) large liptinite (algal) body and thinner spores under blue-light excitation, No. 3 seam.

in modern mires arises due to hydrological modification of the mire surface and is therefore a measure of groundwater influence, albeit an indirect one. The role of desmocollinite in a measure of groundwater influence is therefore problematic.

Calder *et al.* (1991) originally defined the GWI as gelocollinite + corpocollinite + desmocollinite (biochemically and physically degraded tissues) + mineral matter: telinite + telocollinite (structured and partially structured tissues), although they discussed the exclusion of corpocollinite because of uncertain origins and desmocollinite because of cellulose affinity.

4.9.3 Macerals as Indicators of Vegetation

Theoretically, mires can be broadly subdivided on the basis of vegetation type by contrasting macerals of forest affinity with those of herbaceous and marginal aquatic affinity, although the accurate reconstruction of vegetational cover may exceed the capability of conventional maceral analysis. More accurate paleoecological reconstruction may be possible through phyteral analysis (Winston, 1989). Phyterals (from the Greek *phyton*, plant) are plant parts in coal recognizable under the microscope and for which the parent plant affinity is known. Lignin-rich vegetation of a forest habitat is represented by telocollinite, fusinite and semifusinite; suberinite and resinite are associated macerals. Although gelocollinite and corpocollinite are assumed to be derived largely from lignin and tannins, respectively, their origin is less clear than the origin of telinite and telocollinite, for instance, hence their exclusion in this version of a vegetation index (see below). A paleomire plot for the inner mire region that excludes gelocollinite and corpocollinite from the vegetation index (Figure 4.9a) is compared with one including the two maceral types in Figure 4.9b. Herbaceous, cellulose-rich vegetation is represented in part by desmocollinite, inertodetrinite and liptodetrinite (Teichmüller, 1989; Mukhopadhyay, 1989). Truly aquatic conditions are indicated by alginite, representing planktonic algae. No higher plants (macrophytes) from the Late Carboniferous have thus far been identified as truly aquatic (Collinson and Scott, 1987), and thus the term "marginal aquatic" is invoked to represent the truly aquatic algae plus those macerals which preferentially accumulated in the limno-telmatic zone, irrespective of their site of origin (e.g. sporinite). Sporinite and cutinite are included in this study because of their similar distribution to other marginal aqautic macerals within the No. 3 seam. The resulting measure of vegetation type, referred to as the **vegetation index (VI)**, can be expressed as:

$$\frac{\text{telinite} + \text{telocollinite} + \text{fusinite} + \text{semifusinite} + \text{suberinite} + \text{resinite}}{\text{desmocollinite} + \text{inertodetrinite} + \text{alginite} + \text{liptodetrinite} + \text{sporinite} + \text{cutinite}}$$

The vegetation index is similar to the "tissue preservation index" (TPI) of Diessel (1986), which was given as:

$$\frac{\text{telinite + telocollinite + semifusinite + fusinite}}{\text{desmocollinite + macrinite + inertodetrinite}}$$

Named the tissue preservation index (TPI), it addresses change in tissue preservation only indirectly as a function of the type of plant tissue (woody vs. herbaceous) entering the system. The TPI of Diessel (with the removal of macrinite) in contrasting lignin-rich, arboreous with cellulose-rich, presumably herbaceous tissues may reflect a mire's evolution toward a raised/domed condition (Figure 4.1), although such a use was not suggested. The main departure in this study from Diessel's TPI is the inclusion of vegetation and plant matter of aquatic affinity. A similar maceral ratio was developed by Mukhopadhyay (1989) for Texas Tertiary coals to characterize various wetland settings (swamp, swamp-marsh, marsh, aquatic).

4.9.4 Mire paleoenvironment diagram

The mire paleoenvironment diagram (Figure 4.8a) is a maceral-based method by which an ancient peat-forming ecosystem can be modelled using parameters similar to those employed in classification of modern mires, as described earlier. The method further develops the concept introduced by Diessel (1982, 1986), *but differs in theory*, primarily in the definition of a measure of groundwater influence, the fundamental criterion for mire classification, and the abandonment of the long-standing use of relative Eh levels as deduced through ratios of fusinitized vs. vitrinitized tissues.

The mire types identified on the diagram (Figure 4.8a) are the hydrological types swamp, fen and bog, with vegetational subtypes swamp forest and bog forest. The limnic environment, site of sapropelic accumulation, is also represented. Wetland environments prone to siliciclastic inundation result in the formation of impure coal lithotypes (shaly coal, coaly shale) and carbonaceous mudrock. The high ash content of such deposits places them outside the definition of peat, hence their environments are not truly mire types. The terms 'inundated marsh' and 'inundated forest' were thus employed by Calder *et al.* for coals exhibiting a GWI in excess of 5. The mire types from top to bottom in the mire paleoenvironment diagram (Figure 4.8a) represent an evolution from rheotrophic to ombrotrophic conditions, and planar to domed morphology accompanied by a change from

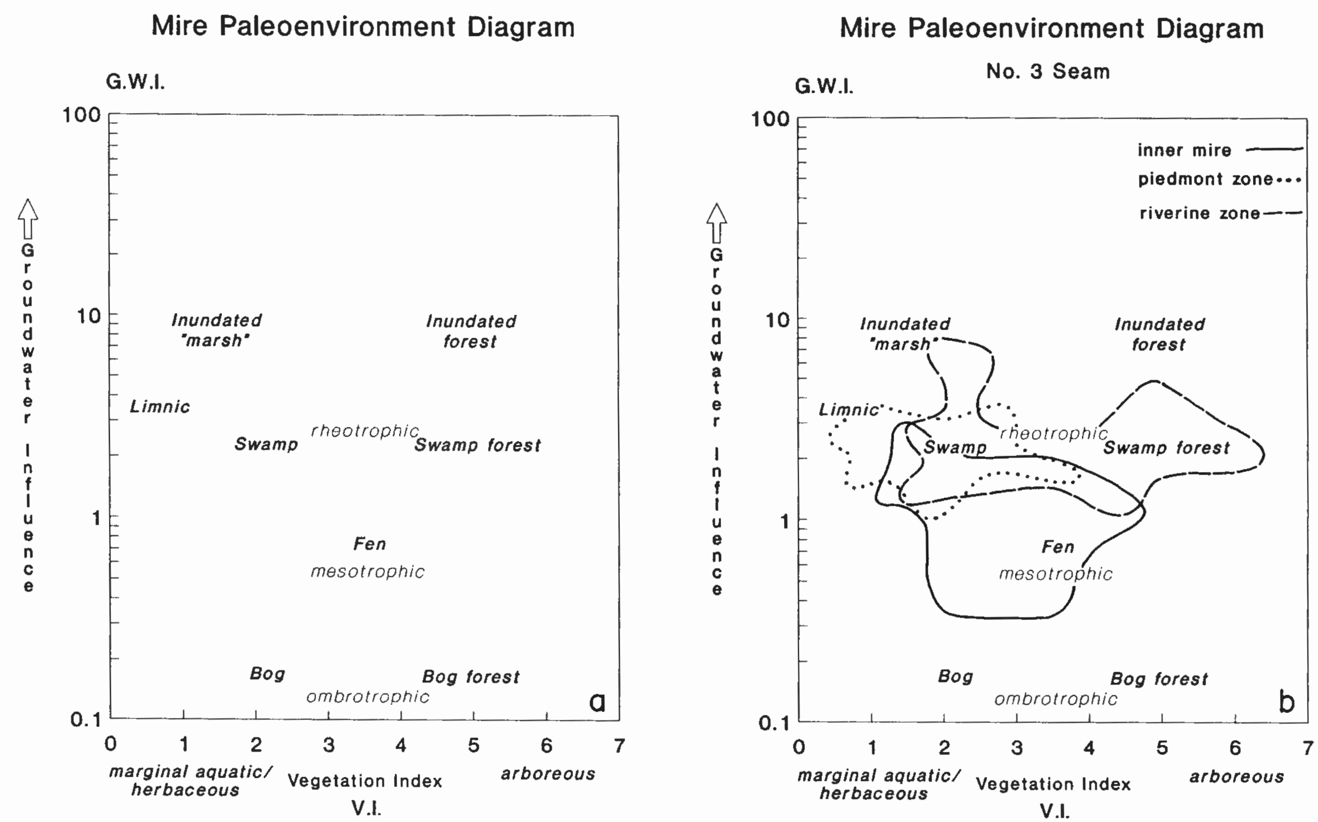


Figure 4.8 a) The mire paleoenvironment diagram (modified from Calder *et al.*, 1991). b) Comparison of paleomire types that typify the piedmont, inner mire and riverine zones of the No. 3 seam.

eutrophic to oligotrophic nutrient status (see also Figure 4.1), but index values are somewhat arbitrary, indicating *trends* rather than absolute limits. The GWI value below which ombrotrophic conditions prevailed will be more accurately constrained as other seams are evaluated.

The vegetation index, described previously, was reluctantly used in an attempt to provide the required template of vegetation type in order to achieve a wholly-maceral-based method of ancient mire habitat that honoured all major templates of modern mire classification. As discussed above, however, at our present level of understanding of botanical and ecological affinities of macerals, conventional maceral analysis is a poor avenue for reconstruction of vegetation type. Trends in vegetation type are to a large degree a function of groundwater influence, therefore groundwater influence and vegetation indices will be very much inter-related.

4.10 PALEOMIRE RECONSTRUCTION OF THE NO. 3 SEAM

According to the maceral-based indices of groundwater influence and vegetation, distinct mire types developed within the three areal zones (piedmont, inner mire and riverine) of the peat-forming system (Figure 4.8b), the sole type common to all three zones being the swamp environment. The inner zone of the ancestral No. 3 mire records a systematic evolutionary sequence of mire types throughout its development (Figure 4.9a,b), inferred to represent a hydroseral succession (Tallis, 1983). The lower third of the seam developed under highly rheotrophic conditions and inundated marsh/limnic to swamp environments. The mire may have subsequently evolved to more forested conditions but felt the effects within the inner mire of the proximity of a channel-belt which occupied the northern, riverine zone of the mire adjacent to the basin axis. The GWI indicates that the surface of the inner mire continued to rise relative to groundwater level with the development of fen and possibly bog and ultimately bog forest mire types, following abandonment of the riverine zone by this fluvial system in favour of a more northerly axial position. Increases in ash, sulphur and miospore diversity and a trend to duller lithotypes within the upper third of the seam are contrary evidence of continued doming, and may suggest deflation of the mire surface. (See section on processes of mire development).

Both riverine and piedmont margins of the mire developed under rheotrophic conditions. The piedmont zone records a weakly developed hydroseral succession of mire types but did not evolve beyond the swamp stage; inundated forest characterized this margin during much of its history. The two leaves (benches) of the No. 3 seam in the riverine zone developed from different mire types: the

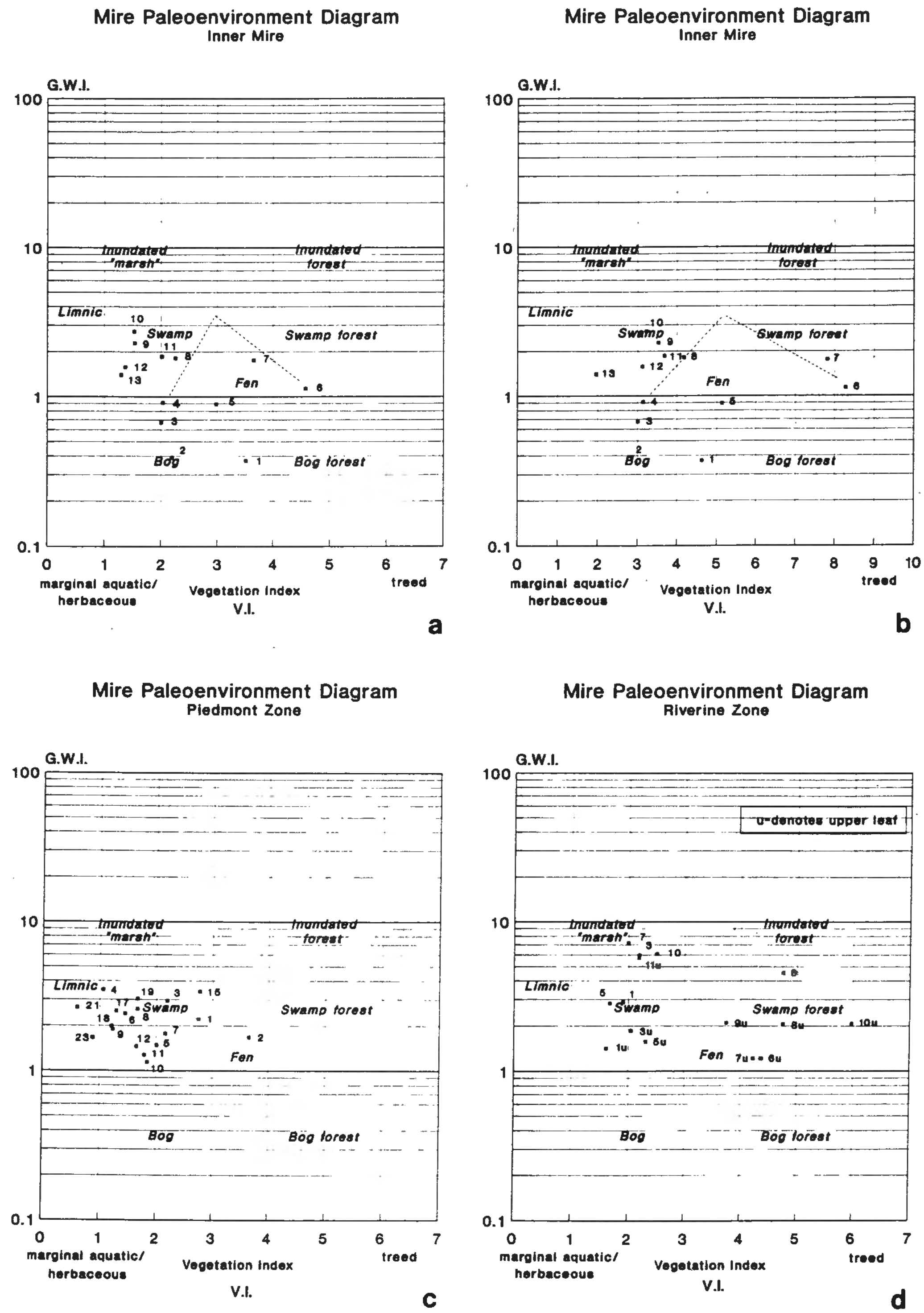


Figure 4.9 a) Evolution of paleomire types within the inner mire zone, No. 3 seam, as suggested by maceral analysis. Dashed line indicates inferred actual pathway in section of lost core that comprises a thin mudrock parting. b) Paleomire evolution of the inner mire including gelocollinite in the Vegetation Index (assuming derivation from lignin, hence arboreous vegetation). c) Paleomire evolution of the piedmont zone, No. 3 seam. d) Paleomire evolution of the riverine zone, No. 3 seam. Sample nos. correspond to those in Figures 4.5, 4.6 and 4.7.

lower leaf accumulated within an inundated marsh environment, whereas the upper leaf evolved to swamp and fen. The different petrographic composition and paleomire types of the two leaves can be partly attributed to the mesotrophic tendency of the system and the diastem of peat development during fluvial occupation. The contrast is likely due in part to relief inherited from the different compaction of sediment around the entombed multistorey sandstone body.

The ancestral mire of the No. 3 seam therefore was predominantly rheotrophic, but exhibited a mesotrophic tendency to less groundwater-influenced conditions.

4.10.1 Ash and Sulphur Distribution

Coal seams of the Springhill coalfield exhibit a typical areal distribution of ash and sulphur content, with the margins enriched in both components relative to the inner areas. The No. 3 seam is typical of the Springhill coals in that the inner or central region has a total sulphur content (by weight) of 1.6%, which increases to greater than 3% (total seam) at seam margins. The participation of sulphate, pyritic and organic sulphur differs between the inner mire and the relatively sulphur-enriched margins. Within the inner mire, organic sulphur is the greatest contributor and sulphates contribute the least, whereas in both marginal zones pyritic sulphur is the most abundant form. The ash content of the seam increases sharply and steadily away from the inner zone (10.1 % by weight, SH81), especially toward the piedmont margin. Within the inner zone, the content of ash and sulphur is greatest at the seam base, and lowest near mid-seam (Table 4.7).

Upwardly decreasing pH level within the inner mire, inferred from the increasing ratio of structured tissues (telinite, telocollinite) to strongly gelified tissues (structureless collinite, gelocollinite *sensu latu*), accompanied raising of the mire surface and diminishing influence of groundwater. At the piedmont and riverine margins and during the early formation of the inner mire, alkaline ions are inferred to have been supplied by eroding Early Carboniferous limestones (Bell, 1927; Gibling *et al.* 1989) or by fan discharge and fault-fed springs. Sulphate-reducing bacteria would have thrived under the resulting conditions of elevated pH (Casagrande, 1987; Cecil *et al.*, 1979; Altschuler *et al.*, 1983). The enrichment of sulphur, (particularly as pyrite), in areas of rheotrophic influence would be further enhanced by the flocculation of ferrous iron-supplying clays carried into the mire during flooding. Rapid flocculation occurs when clays are introduced to humic acids (Staub and Cohen, 1979; Cecil *et al.*, 1979). As the mire evolved, the inner reaches would have become more acidic and less hospitable to bacteria, while becoming evermore insulated from the introduction of ferrous iron

Piedmont Zone			Inner Mire			Riverine Zone		
Sample	Ash	Sulphur	Sample	Ash	Sulphur	Sample	Ash	Sulphur
SH85-1T	20.69	2.57	SH81-1T	8.58	1.54	SH72-1+2UT	85.35	2.32
SH85-2T	91.30	0.28	SH81-2T	11.70	1.44	SH72-3UT	53.42	5.26
SH85-3T	31.42	1.04	SH81-3T	10.90	1.56	SH72-4UT	91.99	1.94
SH85-4T	76.67	0.60	SH81-4T	8.06	2.06	SH72-5UT	28.87	4.77
SH85-5T	18.46	2.99	SH81-5T	6.18	1.07	SH72-6UT	10.57	4.36
SH85-6T	23.83	3.00	SH81-6T	3.76	0.88	SH72-7UT	18.62	4.32
SH85-7T	25.78	2.33	SH81-7T	4.65	0.87	SH72-8UT	7.90	2.31
SH85-8T	57.53	2.03	SH81-8T	10.08	1.48	SH72-9UT	18.16	2.69
SH85-9T	24.69	4.04	SH81-9T	10.78	1.62	SH72-10UT	8.09	2.93
SH85-10T	46.96	2.00	SH81-10T	13.82	1.57	SH72-11UT	14.52	2.79
SH85-11T	31.11	6.35	SH81-11T	10.25	2.79			
SH85-12T	44.77	20.70	SH81-12T	7.35	2.69	SH72-1LT	51.20	3.61
SH85-13T	60.75	2.26	SH81-13T	25.61	2.38	SH72-2LT	91.93	0.32
SH85-14T	78.70	1.95				SH72-3LT	48.49	7.70
SH85-15T	71.55	1.24				SH72-4LT	92.28	0.65
SH85-16T	56.50	3.81				SH72-5LT	79.45	1.90
SH85-17T	29.82	6.00				SH72-6LT	30.61	7.33
SH85-18T	63.21	1.79				SH72-7LT	85.25	1.14
SH85-19T	40.86	4.89				SH72-8LT	35.17	7.17
SH85-20T	91.04	0.03				SH72-9LT	92.64	0.35
SH85-21T	74.00	1.47				SH72-10LT	32.18	5.40
SH85-22T	90.13	0.29				SH72-11LT	34.27	6.43
SH85-23T	71.64	2.03				SH72-12LT	87.57	1.34
SH85-24T	43.43	4.93				SH72-13LT	71.49	3.98
SH85-25T	53.25	6.86						

Table 4.7 Ash and sulphur content (by weight percent) of samples from the three zones of the No. 3 seam (SH81-1T to 13T and SH85-1T to 11T a.d.b., all others d.b.)

due to increasing concentration of humic acids. The distribution of sulphur may have been controlled by the availability of iron, not only from clays, but also as iron sulphate from oxidation of sulphide-bearing rocks. Such a mechanism may partly explain the enrichment of sulphur at the piedmont margin, which was less likely than the riverine zone to have been influenced by reworked evaporites.

The resulting distribution of ash and sulphur within the brightening upward, lower two-thirds of the inner mire of the No. 3 seam is similar to that of the forest swamp peat of the Baram River area of tropical Sarawak (Figure 4.10) described by Esterle *et al.* (1989) and Cameron *et al.* (1989). The latter authors described a minerotrophic (rheotrophic) to ombrotrophic evolution for these tropical mires. The higher ash content and less pronounced upward decline in sulphur content of the No. 3 seam throughout, however, suggests that the paleomire was more strongly influenced by groundwater than are the tropical Indonesia mires, although peats in general may be poorer in mineral matter than a descendant coal (Cecil *et al.*, 1982).

4.10.2 Comparison with the method of Diessel (1986)

Tissue Preservation and Gelification Indices of Diessel (1986) for the three areal zones of the No. 3 seam are plotted in Figures 4.11a-d for purposes of comparison with the method of the author (Figures 4.8b, 4.9). The petrographic trends of these three seam sections are reflected in the diagrams, but, not surprisingly given the differences in theory behind the two methods, the interpretation of mire type (Figure 4.11a: "coal facies diagram" of Diessel, 1986) differs from that obtained from the mire paleoenvironment diagram of this study. Two terms used by Diessel to describe mire type are not used in the mire paleoenvironment diagram. *Marsh*, in the usage of Moore (1987) implies a non peat-forming herbaceous wetland (Table 4.1) The term *dry forest swamp*, in the opinion of the writer, is paradoxical. A swamp (Table 4.1) is a perennially flooded, flow-fed mire, thus "wet" by definition (Moore, 1987). The coal facies diagram of Diessel does not specifically address the trophic status of mire ecosystems, however, it may be advisable to substitute bog for dry forest swamp. The two methods both indicate that the inner zone of No. 3 seam differed in mire type from the margins and that there is some commonality in all three zones (cf. Figures 3.8b and 4.11a). The method of Diessel indicates that paleomire types in the riverine and piedmont zones were similar and that these marginal areas had evolved to "drier" mire types than did the inner zone (Figure 4.1b). This contradicts the interpretation based on the mire paleoenvironment diagram. All three zones are shown to exhibit distinct mire types and the inner zone, rather than the piedmont and riverine zones,

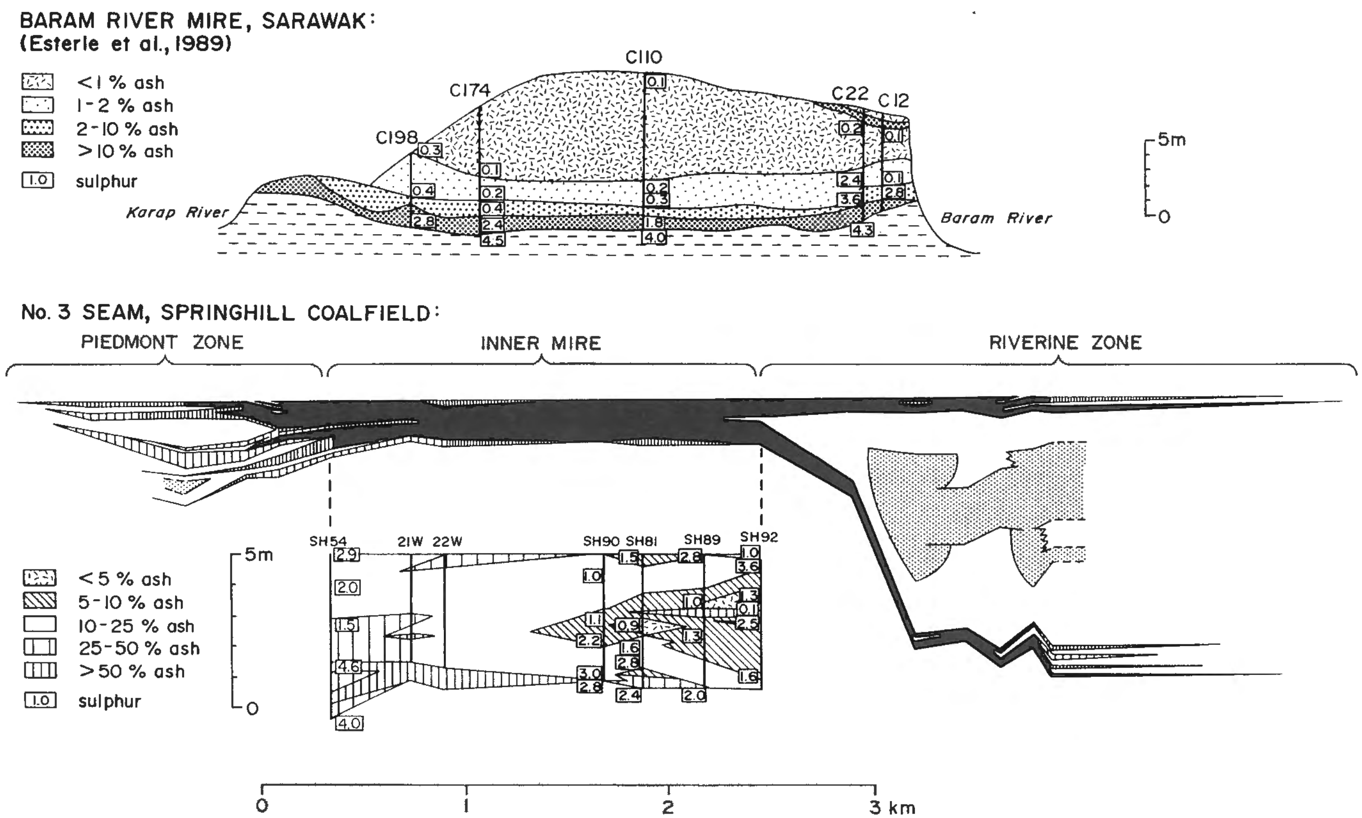


Figure 4.10 Comparison of ash and sulphur distribution within the Baram River mire of Sarawak (after Esterle *et al.*, 1989) and the inner mire, No. 3 seam.

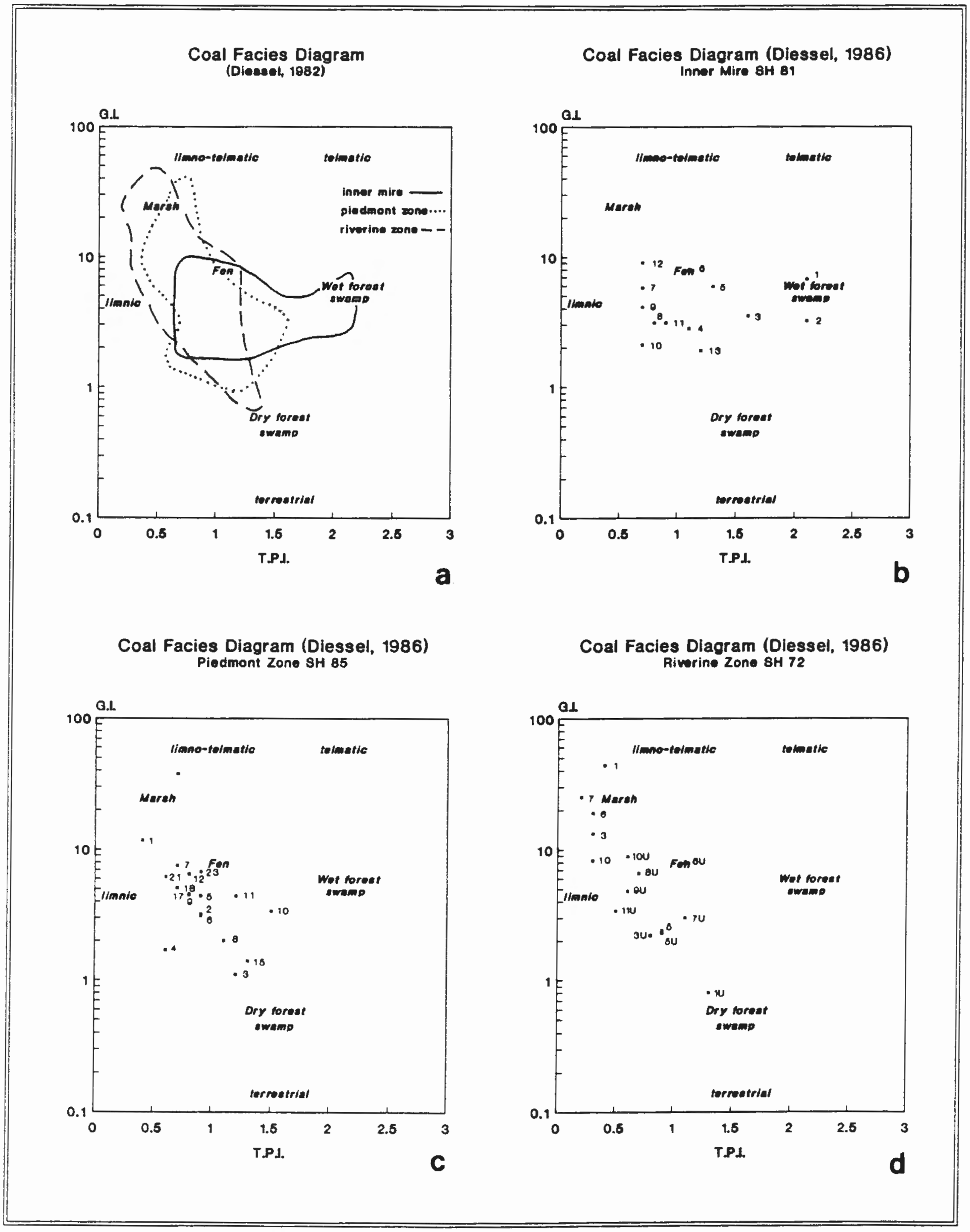


Figure 4.11 Paleomire types for the No. 3 seam suggested by the maceral indices of Diessel (1986): a) comparison of piedmont, inner and riverine zones; b) inner mire zone; c) piedmont zone; d) riverine zone.

evolved to the least groundwater influenced condition (Figure 4.8b). These differences of interpretation can be attributed in large part to the use of inertinite as an indicator of dry conditions.

The inner mire section (SH81, Figure 4.2, 4.3), according to the interpretation of Diessel (Figure 4.11b), represents a fen which evolved in late stages (samples 3 to 1T) to wet forest swamp. The basal sample (13T) which is rich in fusinite (10.6% t.c.) and semifusinite (5.2% t.c.),but also mineral matter (16.4%), paradoxically represents the most terrestrial conditions within the mire in marked contrast to the interpretation of inundated forest to rheotrophic swamp in the mire paleoenvironment diagram of this study (Figure 4.8a). The major variation is in TPI rather than GI; TPI is the highest of the three areal sections, indicating that the relative contribution of lignin vs. cellulose was greatest (Figure 4.11a) in the inner mire. According to Diessel (1986), this would also indicate greatest tree density in the inner mire, but it may be impossible to deduce the density of trees by such a method (DiMichele, written communication, 1989).

The coal facies diagram of the riverine section (SH72, Figure 4.11d) illustrates the contrasting petrographic character of the two seam leaves. The lower leaf would be interpreted in this case as "marsh" to fen with a strongly limnic character, whereas the upper leaf plot indicates predominantly fen, to "dry forest swamp". The lower leaf has a TPI of < .5 (except 5LT which has a TPI of 0.9). The upper leaf has a TPI < 1.3. The so-called Gelification Index exhibits strong variation, reflecting the inertinite content and oxidation of tissues, which is 2-3 times greater in the upper than lower leaf.

The piedmont section (SH85, Figure 4.11c) plots roughly similar to the riverine section, characterized by fen conditions with periodic shifts to a dry and less commonly wet, forest swamp environment, ending in a limno-telmatic "marsh". Here, as in the riverine section, the strongest variation is within the G.I.; T.P.I. does not exceed 1.5, which is substantially lower than in the inner mire.

Diessel (1986, his Figure 1) plotted maceral indices for Australian coals from a variety of depositional settings, and specific despositional environments were shown to plot in similar positions. While this appears to be of use in interpreting the environments of *total seam sections for Australian coals*, caution is suggested in transferring these paleoenvironmental interpretations to coals elsewhere. The wide range of paleoenvironments suggested for the piedmont, inner mire and riverine zones of the No. 3 seam when plotted on the 'coal facies diagram' illustrates this point. The inner mire section of the No. 3 seam plots in a position characteristic of Australian upper delta plain/alluvial valley coals. The lower leaf of the seam in the riverine zone plots in a position similar to the lower delta plain coals of Diessel (1986), whereas the paleoenvironment suggested for samples from the upper leaf in the same zone is predominantly upper delta plain/alluvial valley to piedmont. A similar paleogeographic setting is suggested for the piedmont zone section of the No. 3 seam which similarly plots in a position characteristic of Australian coals from interpreted upper delta plain/alluvial valley, to piedmont environments.

4.11 PALYNOMORPH ANALYSIS

In this section, palynomorphs within the No. 3 seam will be investigated to determine whether the miospore record reflects the interpreted temporal and zonal development of the paleomire. Two of the attributes described by Anderson (1983), i.e. floral change and a reduction in species richness temporally (vertically) and areally (toward the inner mire) can be investigated through miospore analyses, thus providing a basis to assess the evolutionary development of the No. 3 paleomire. Further insight into conditions within the mire can be provided by considering the paleoecology of the parent flora of the miospores.

The floral type and parent plant for miospores identified in this study, (Table 4.8) are given in Smith (1962), Smith and Butterworth (1967), Phillips *et al.* (1985), Bartram (1987), Mahaffy (1988) and Willard (1989a, b). The miospore taxa are grouped by floral type: arboreous lycopsids, herbaceous lycopsids, sphenopsids, tree ferns (Marattiales), small ferns (Filicinae), seed ferns (Pteridosperms), and cordaitean gymnosperms.

4.11.1 Methods

All samples submitted for maceral analysis were halved and tendered to Dr. G. Dolby, Calgary, for detailed palynological determinations (Dolby, 1988a). Samples were first treated with hydrofluoric acid to remove silicates, and subsequently with Schultze solution to extricate spores from the coal. Slides were made of the sieved (+10 and +30 micron) fractions from the concentrate. The +10 micron slides were used for the analyses; the +30 micron slides were subsequently checked for larger forms.

For each sample, an initial 200 counts were made of the total palynomorphs to obtain population percentage. A secondary count of palynomorphs was then made, discounting the genus

Miospore Genus	Paleobotanical Affinity	Authority	Citing Reference
	Arboreous Lycopsids		
Lycospora	Lepidodendron	Chaloner, 1953	Smith, 1962
	Lepidophloios	Andrews & Pannel, 1942	Smith, 1962
	Paralycopodites (Anabathra)	DiMichele, 1980	Bartram, 1987
L. pusilla	Lepidodendron hickii	Willard, 1989	-
L. orbicula	Paralycopodites (Anabathra)	Willard (pers. comm.)	-
L. pellucida	Lepidophloios harcourtii	Willard, 1989	-
L. granulata	Lepidophloios hallii	Willard, 1989	-
Crassispora	Sigillaria	Chaloner, 1953	Bartram, 1987
Granisporites,	Diaphorodendron	DiMichele, 1985	Bartram, 1987
Apiculatisporis			Phillips et al. 1985
Densosporites cf. sphaerotriangularis	Sporangiostrobus	Leisman, 1970	Bartram, 1987
	Herbaceous Lycopsids		
Densosporites cf. loricatus	Porostrobus	Chaloner, 1958	Bartram, 1987
Endosporites	Chaloneria	Pigg & Rothwell, 1983	Bartram, 1987
Cirratriradites cf. saturnii	Sellaginellites	Hoskins & Abbott, 1956	Bartram, 1987
			Smith & Butterworth, 1967
Cingulizonates			
Cristatisporites			Phillips et al., 1985
Radiizonates			Phillips et al., 1985
	Ferns		
Granulatisporites	? Pteridosperm	Knox, 1938	Smith, 1962
Planisporites	? Pteridosperm	Kidston, 1923-25	Smith, 1962
Laevigatosporites (species < 40 microns)	tree fern (Marattiales)		Mahaffy, 1988
Punctatosporites	tree fern (Marattiales)	Mamay, 1950; Remy & Remy, 1957	Smith, 1962
Cyclogranisporites	var. ferns	Remy & Remy, 1955	Smith, 1962
Leiotriletes	fern (Filicales)	Knox, 1938; Mamay, 1950; Remy & Remy, 1960	Smith, 1962
Lophotriletes	fern	Remy & Remy, 1957	Smith, 1962
Raistrickia	fern (Filicales)	Radforth, 1938; Mamay, 1950; Remy & Remy, 1955	Smith, 1962
Savritrisporites	Senftenbergia (Filicales)		Smith & Butterworth, 1967
Reticulatisporites	fern (Filicales; Coenopteridales)		Phillips et al., 1979
	Sphenopsids		
Calamospora	Calamites	Hartung, 1933	Smith, 1962
	Sphenophyll	Remy & Remy, 1955	Smith, 1962
Laevigatosporites (spp. > 40 microns)	Calamites	Reed, 1938; Andrews & Mamay, 1951; Remy, 1960	Smith, 1962
Vestispora			Mahaffy, 1988
	Seed Plants (Gymnosperms)		
Florinites	Cordaites	Florin, 1936, 1938-40	Smith, 1962
Schopfipollenites	Medullosa		Phillips et al., 1985
indet. pollen			

Table 4.8 Paleobotanical affinities of miospore genera reported from the No. 3 seam. For bibliographic information, see citing reference.

Lycospora, so as to obtain a record of as many species as possible. Due to low spore abundance in many samples, 200 counts were not always possible, necessitating a conversion of the data from counts to percentage.

4.11.2 Results of Palynological Analysis

The results of palynomorph analysis can be summarized as follows:

1) Miospores of the genus *Lycospora* are predominant in all but a few samples in the three areal zones of the No. 3 seam (Figures 4.12, 4.13 and 4.14); the two most abundant miospore species with the No. 3 seam are *Lycospora pusilla* and *L. orbicula*.
2) The genera *Calamospora* and sub-dominant *Florinites* are co-dominant with the *Lycospora* in certain samples associated with fluvial deposits of the riverine to inner zones of the mire (Figures 4.12 and 4.14).
3) There is a striking paucity of miospores of both herbaceous lycopsid and fern affinity (Table 4.8).
4) Species diversity is greater at the piedmont margin (61 species) than in the riverine zone or inner mire sections (44 species each). An overall upward trend of decreasing diversity from 27 through 9 species, followed by a partial recovery to 17 species, is evident within the inner mire (Figure 4.12).
5) Species that occur only in one or two mire zones (Table 4.9) are low in abundance and generally associated with siliciclastic partings.
6) The miospore analyses record a marked vertical change in relative proportions of *Lycospora* within the inner mire section (Figure 4.12). Two and possibly three palynomorph assemblages are evident: 1) an assemblage co-dominated by *Lycospora pusilla* and *L. orbicula* but with *L. orbicula* becoming dominant through time in a large-scale cycle through the lower two-thirds of the seam (SH81-13 to 5T); 2) a second assemblage marked by a sharp decline in L. orbicula and concomitant increase in *L. pusilla* (SH81-4 to 3T); and 3) a tentative assemblage marked by the presence of *L. pellucida*, but dominated by *L. pusilla* (SH81-2 to 1T).
7) In marginal zones of the seam (Figures 4.13 and 4.14) further palynomorph assemblages are evident: 4) an assemblage wherein *L. pellucida* was a near equal subdominant with *L. orbicula*; and 5) an assemblage marked by the presence of *L. granulata*, the least abundant species of *Lycospora*, occurring primarily in the basal section of the seam at the riverine margin.

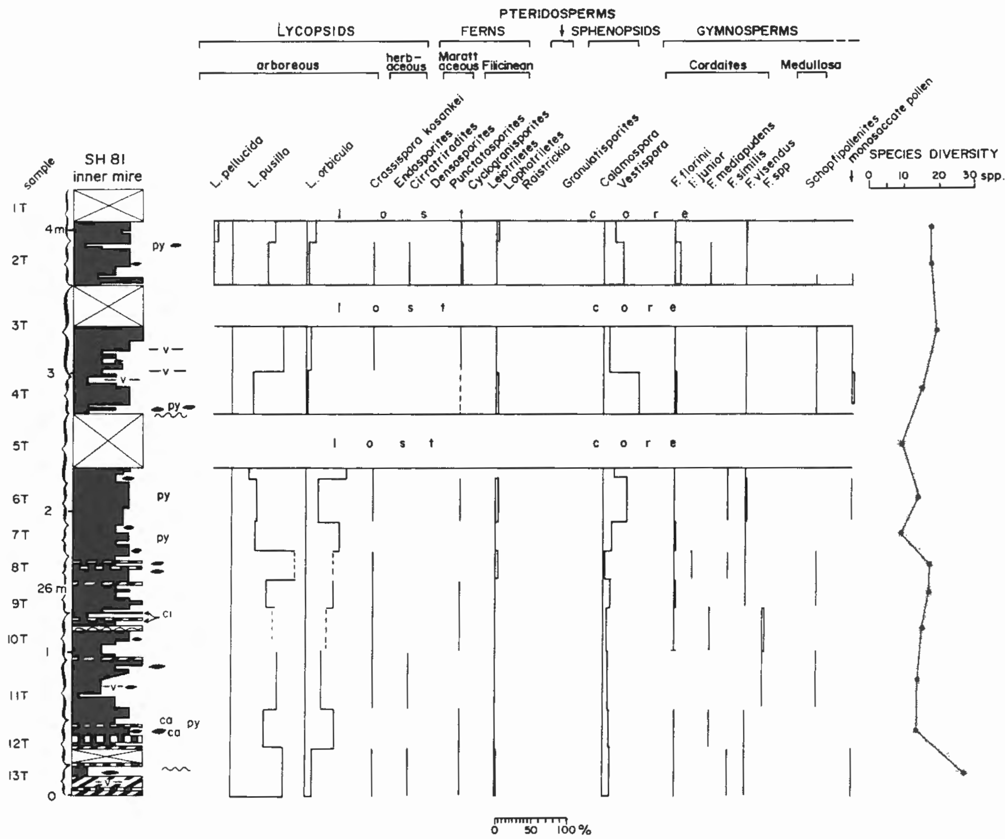


Figure 4.12 Palynology profile, inner mire, No. 3 seam (drillhole SH81).

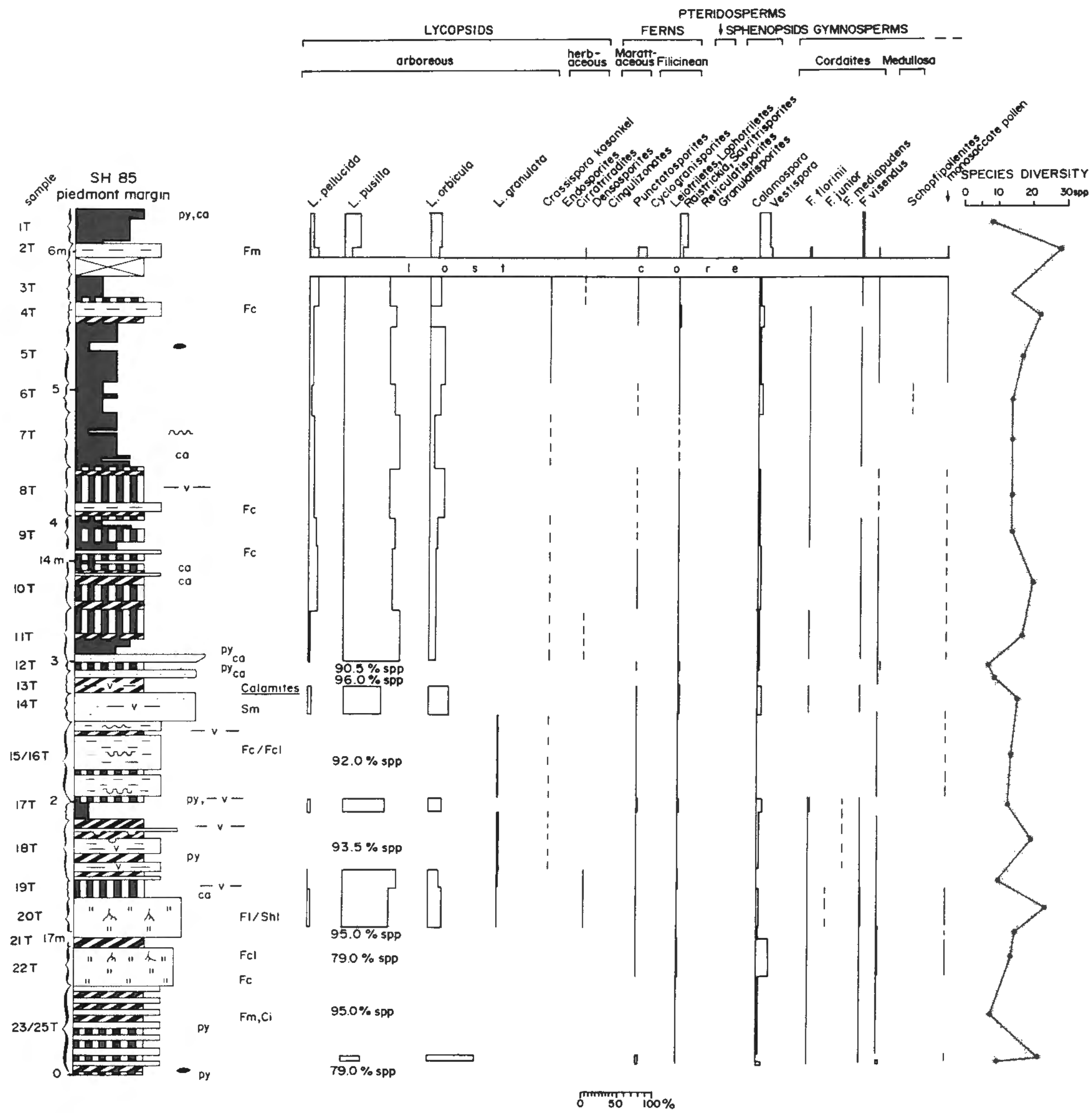


Figure 4.13 Palynology profile, piedmont zone, No. 3 seam (drillhole SH85).

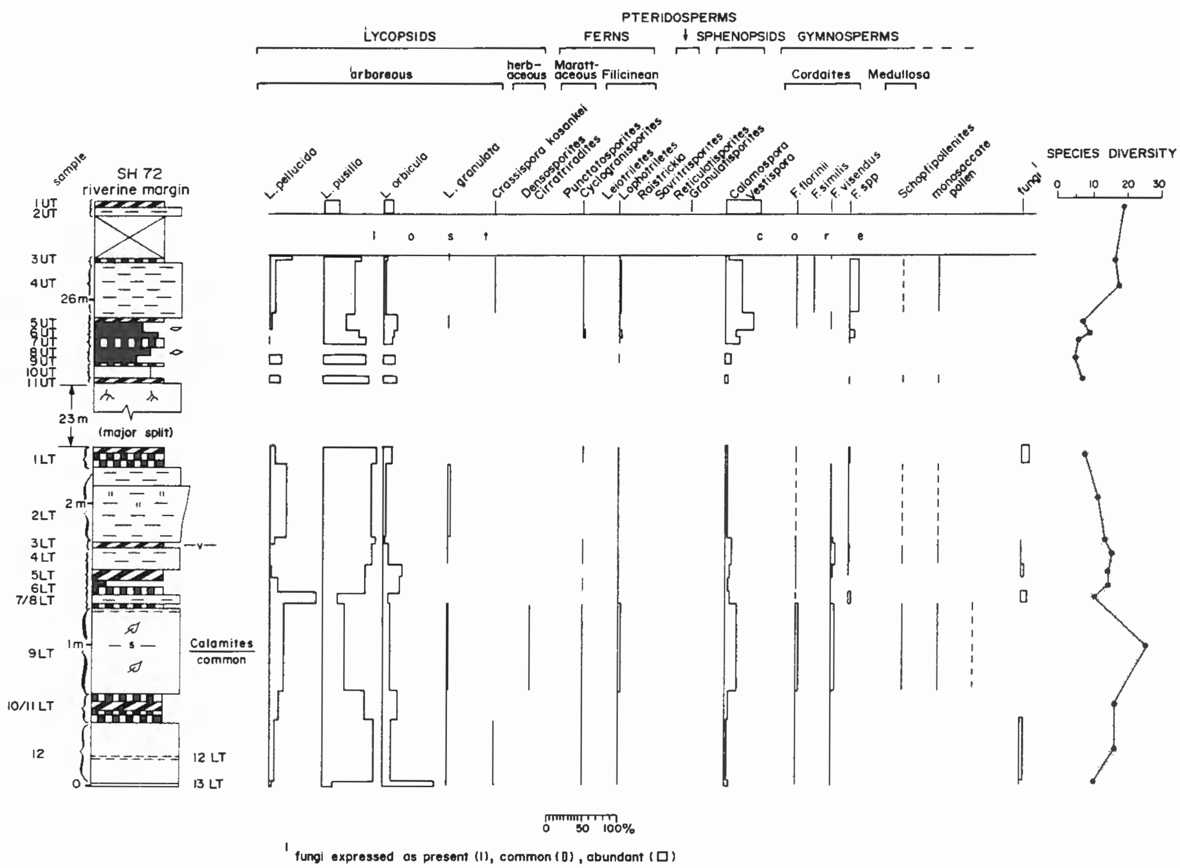


Figure 4.14 Palynology profile, riverine zone, No. 3 seam (drillhole SH72).

Miospore species unique to one mire zone:

PIEDMONT	INNER MIRE	RIVERINE
Acanthotriletes spp.	*Dictyotriletes muricatus*	Bisaccate pollen
Cingulizonates spp.	*Playfordiaspora 38/16*	*Lophotriletes mosaicus*
Cristatisporites spp.	*Playfordiaspora 55/16*	
Microret. cf. nobilis	*Tantillus triquetrive*	
Plicatipollenites "minutus"		
Radiizonates spp.		
Wilsonites delicatus		
Schulzopora spp.		

Species restricted to two zones:

PIEDMONT - INNER MIRE	RIVERINE-INNER MIRE	PIEDMONT-RIVERINE MARGINS
Florinites mediapudens	*Camptotriletes sp.*	*Apiculatisporis spp.*
Grumosisporites vario.	*Florinites similis*	*Auroraspora solisorta*
Knoxisporites triradritus	*Playfordiaspora sp.*	*Cananoropollis janakii*
Raistrickia superba	*Guthoerlisporites sp.*	*Cirratriradites spp.*
Retusotriletes spp.		*Granulatisporites cf. pallidus*
Mooreisporites fustis		*Plicatipollinites spp.*
Endosporites globiformis		*Punctatisporites spp.*
Cirratriradites saturni		*Raistrickia lacerata*
		Rugospora gracilirugosa
		Savritrisporites asperatus
		Walzispora planiangulata

Table 4.9 Miospore species recorded from one or two zones only of the No. 3 seam.

8) Fungal spores occur in abundance within the lower leaf (bench) of the seam in the riverine zone.

4.12 INTERPRETATION

The use of miospores in paleoecological study is limited by constraints similar to those that limit the use of macerals. The main constraints are: 1) macrofloral (i.e. parent plant) affinity; 2) effects of transport; and 3) variable spore production leading to over-representation by different species. In the evaluation of mire ecology of the No. 3 seam, an additional factor is the corroded state of some miospores particularly within the piedmont zone (Dolby, 1988a),inferred to be due to bacterial activity, high pH and rheotrophic hydrology. Not all spores are equally likely to survive oxidation; therefore, there is a potentially strong taphonomic bias.

4.12.1 Miospore Diversity

An upward (temporal) and inward (areal) decrease in species diversity was cited by Anderson (1983) as one of the diagnostic features of the floral zonations (catena) of tropical Malaysian and Indonesian mires. Within the No. 3 seam, miospore species diversity is generally greater at seam margins and within the basal inner mire zone (Figures 4.12-4.14). The increase in species diversity in these areas can be linked to rheotrophic hydrology, because rheotrophic areas of the mire, which experienced water input from neighbouring deposystems should contain a higher proportion of externally-derived spores and pollen (Fulton, 1987). These externally derived palynomorphs will occur in greatest abundance within siliciclastic partings (Marshall and Smith, 1965) introduced to the mire periphery during floods if the partings were impervious enough to protect the spores from decay.

The relationship between increased groundwater flow or rheotrophic conditions and increased miospore diversity is not simply a function of enhanced transport. Increased environmental stress such as waterlogged conditions, relatively low nutrient supply and decreased pH would have had the effect of lowering species diversity within the inner mire. In areas at the mire periphery and beyond, where nutrient supply was greater, species diversity would have been enhanced.

Within the inner mire, species diversity is greatest at the seam base (n = 27), decreasing steadily upward (n = 9) to mid-seam, but then experiencing a partial recovery (to 17) at the top. The lower trend is the one to be expected from a tropical forested mire such as those of tropical Indonesia

(Anderson, 1983). The reversal in the upper half is problematic, since maceral data suggest a continued raising of the mire surface. This, and other trends that appear to contradict evidence of doming, are discussed further in the sections on processes of mire development. The significance of these trends in species diversity must be weighed against the absence of replicate sample analysis and the very low abundance of many species.

4.12.2 General Plant Community Composition

Macrofloral affinities of miospores (Table 4.8) can be used to provide a reconstruction of the peat-forming plant communities (Scott, 1978), although there are inherent biases due to modes of dispersal and preservation (Phillips *et al.*, 1985). The miospore genus *Lycospora* is dominant in all but a few samples from the three areal zones of the No. 3 seam. Even though *Lycospora*-producing lycopsids are one of the most over-represented in the miospore record (Phillips *et al.*, 1985), the overwhelming abundance of *Lycospora* (as high as 98.5 % in an individual sample from the No. 3 seam) nonetheless suggests that the arboreous lycopsids were the dominant plants throughout the mire. *Sigillaria*, an arboreous lycopsid represented by the miospore *Crassispora kosankei*, was a relatively minor tree not only in the ancestral No. 3 seam but apparently in general throughout the Westphalian (Early and Middle Pennsylvanian) mires of Euramerica (Phillips *et al.*, 1985), although exceptions clearly exist (e.g.Scott, 1978, his Figure 16).

In contrast to the abundance of the arboreous types, miospores attributed to herbaceous lycopsids are all but absent (Figures 4.12 to 4.14). Bartram (1987) has demonstrated, however, that phases of herbaceous lycopsids recorded by megaspores can go undetected in miospore studies due to the exclusion of megaspores from finer sieve fractions; large Medullosan spores may similarly elude detection (DiMichele and Demaris, 1987).

Two floral groups were sub-dominant and rarely dominant within the mire: sphenopsids (calamites) and the cordaitean gymnosperms. In contrast to other mires, true ferns, including both tree (Marattaceous) and herbaceous (Filicinean) types, and the seed ferns (pteridosperms) were minor constituents. *Medullosa*, represented by the prepollen *Schopfipollenites*, was apparently rare, although it is probably under-represented in the miospore record (Phillips *et al.*, 1985).

The abundance of calamitean sphenopsids and cordaitean gymnosperms in relation to the rarely occurring ferns (*sensu lato*) and *Medullosa* may have been partly a function of the adaptive evolution of these plants within mires, as noted by Phillips *et al. (ibid*, p. 95): "in general, cordaites and, to a lesser extent, calamites filled significant habitat gaps in the early Middle Pennsylvanian [circa Westphalian B] until *Psaronius* and *Medullosa* expanded as subdominant plants".

The two most abundant trees and/or prolific producers of miospores within the No. 3 mire were *Lepidodendron hickii (Lycospora pusilla*) and *Paralycopodites*, redefined by Pearson (1986) as *Anabathra (L. orbicula*). In the marginal regions of the mire, *Lepidophloios harcourtii (L. pellucida*) was a near-equal subdominant with *Anabathra*; and in the lower bench of the seam at the riverine margin, was somewhat more abundant than *Anabathra*, assuming equal production and preservation of spores. *Lepidophloios hallii (L. granulata*) was the least abundant of the lycopsid trees, occurring primarily in the early stages of mire development in marginal regions. The general composition of the mire forest, until recently, would have been considered atypical of mire floras of the Late Carboniferous (Phillips *et al.*, 1985; DiMichele *et al.*, 1985). Of particular note is the abundance of *Lepidodendron hickii* since *Lepidodendron* is generally not considered to be "centered in coal swamps" (DiMichele *et al.*, 1985). It is becoming apparent, however, that *Anabathra* - dominated floras are less rare than previously supposed, occurring within the Secor coal of Oklahoma, the Hamlin coal of eastern Kentucky, the Katharina seam of Germany (DiMichele, written comm., 1990) and within zones of the Herrin coal bed (DiMichele and Phillips, 1988) and Low Barnsley seam of Yorkshire (Bartram, 1987).

The paucity of *Lepidophloios hallii*, which was the major tree of some Westphalian D mires (Phillips *et al.*, 1985) may be a function of evolution rather than environment since it did not reach its acme until the Westphalian D when it replaced *L. harcourtii* (DiMichele *et al.*, 1985, their Figure 8.2), however the general paucity of *L. harcourtii* cannot be similarly attributed to evolution. *Lepidophloios* is considered to have been the arboreous lycopsid most adapted to wet, aquatic conditions. Ultimate extinction of *Lepidophloios* has been attributed to a change to a relatively drier, perhaps seasonal, climate (DiMichele *et al.*, 1985; Phillips *et al.*, 1985)

4.12.3 Paleoecology of Arboreous Lycopsids: Reflection of Rheotrophic Conditions

The relative abundance of the arboreous lycopsids within the No. 3 paleomire is dissimilar to the general floral composition of the majority of documented Late Carboniferous mires, many of which, however, are of Westphalian C-D age. This implies that the No. 3 seam developed under paleogeographic, hydrological and chemical conditions that were dissimilar to those of the majority of mires which have been studied in the light of paleoecology, although there may be an evolutionary bias as well. Nonetheless, published interpretations of the paleoecology of flora that dominated the ancient No. 3 mires are in accord with their occurrence in a groundwater-influenced piedmont-riverine setting.

Lepidodendron hickii, the dominant tree of the ecosystem, is considered to have favoured clastic wetlands and "rare nutrient-rich areas, possibly with great freshwater influx" (DiMichele *et al.*, 1985, p. 236) such as the southern, piedmont margin of the Cumberland Basin margin during the Westphalian B. DiMichele and Phillips (1985, p. 236) attributed its occurrence in Lower and Middle Pennsylvanian (Westphalian) peat mires to "unusual conditions, perhaps involving increased clastic supply or seasonal dryness".

Anabathra, an uncommon component of Pennsylvanian peat mires (DiMichele *et al.*, 1985), is thought to have been a "colonizer of disturbed, clastic-rich (transitional) substrates" (DiMichele *et al., ibid*). Current research being conducted by T.L. Phillips of the University of Illinois and by W.A. DiMichele and D.A. Willard at the Smithsonian Institution similarly indicates that *Anabathra* was ecotonal (an ecotone representing a boundary condition between two floral communities), favouring transitional, flood-prone environments. This theory is supported by the author based on the abundance of *Lycospora orbicula* in the lower half of the No. 3 seam, which developed from a swamp habitat under rheotrophic conditions. Its is also consistent with the distribution of *Anabathra* near clastic partings as reported by DiMichele and Phillips (1988). The greater abundance of *L. orbicula* in the inner mire as opposed to the margins and at the high ash base of the inner mire, however, indicates that *Anabathra* was less successful than *Lepidodendron hickii* in the marginal mire areas, perhaps being less tolerant than *Lepidodendron* of siliciclastic input. The decline of *Anabathra (L. orbicula)* in the development of the upper third of the inner mire may indicate an intolerance to lower pH, water level, or nutrient supply. The occurrence of two 1 cm-thick fusain layers at the boundary between the two palynomorph assemblages raises the possibility that this change may have been promoted by wildfire when the *Anabathra* community was becoming increasingly stressed by the hydrological evolution of the mire.

DiMichele *et al.* (1985) interpreted *Anabathra* as an element of a "high diversity, drier site assemblage". The author concurs with the association of *Anabathra (L. orbicula*) and high species diversity (Figures 4.12, 4.14), but evidence from this study conflicts with the interpretation of relatively dry habitat. Apparently this inference arose from the association of *Anabathra* and fusain (Phillips *et al.*, 1985) and the prevailing concept that fusain represents a drier site, which as discussed earlier, may not be valid given its wildfire origin (Scott, 1989).

Lepidophloios was not a major component of the mire vegetation, but its occurrence in the lower parts of the seam in marginal areas supports the adaptation of this flora to a high water table (Phillips, 1979; DiMichele and Phillips, 1985; DiMichele *et al.*, 1985). It is noteworthy that *Lepidophloios hallii* was found to occur in greater abundance within the Herrin coal of Illinois in areas proximal to a contemporaneous paleochannel (DiMichele and Phillips, 1988). A similar, albeit proportionately lesser, increase in abundance of *L. hallii* occurred during development of the lower leaf of the No. 3 seam in the riverine zone of peat accumulation.

4.12.4 Calamitean paleoecology

As noted earlier, the sphenopsids and in particular the calamites, was apparently the only other floral group to periodically become the dominant flora of the mire. The paleoecology of this flora, stem compressions of which are ubiquitous within the coal-bearing strata of the Cumberland Basin, has received considerably less attention in the literature than that of the lycopsids, ferns and pteriodosperms, particularly as a constituent of the "coal swamp" flora.

In studies of Westphalian B Coal Measure floras from northern Britain, Scott (1978, 1979) concluded that calamitean sphenopsids (calamites) occupied lake margin and "delta top" habitats, but also occurred on point bars of river meanders (predominantly in Scott's lithofacies association '2C': carbonaceous shale and coal). Calamitean sphenopsids are thought to have been one of the few "coal age" plants with the ability to propagate and survive in the face of siliciclastic inundation through the growth of adventitious roots (Krassilov, 1972), yet, palynomorphs of the calamitean sphenopsids are less common in regions of the No. 3 mire which formed under the greatest influence of siliciclastic-carrying groundwater, i.e. the piedmont zone (Figure 4.13) and the early deposits of the inner mire (Figure 4.12).

The two main areas within the No. 3 seam in which calamites became dominant or attained a near equal status with the arboreous lycopsids are the upper 2.35m in the inner mire (SH81; samples 6T, 4T and 2T, Figure 4.12) and the upper leaf in the riverine zone (SH72; samples 5UT, 4UT and

1/2 UT, Figure 4.14). The samples from the riverine zone which are rich in *Calamospora* comprise in whole or in part carbonaceous mudrock (lithofacies Fc), impure coal (Ci) and in sample 5UT only, duroclarain (Cdc) with a trace of fusain (Cf). Such an association is in keeping with the marginal lake habitat determined for calamites by Scott (1978, 1979). In contrast, certain mudrock lithofacies, particularly those occurring in the lower leaf, are poor in *Calamospora*. The miospore genus is particularly rare (3.0 percent) in a coarsening-upward carbonaceous mudstone (Fc to Fm; sample 2 LT) which most probably represents a distal crevasse splay/sheetflood deposit. On the other hand, an underlying carbonaceous mudrock (sample 9LT) of uniform grain-size containing diffuse quasi-laminar siderite and common *Calamites* compressions is relatively rich in *Calamospora* (16.5 percent). The relative quiescence of the floodbasin setting may therefore control, at least in part, abundance of *Calamospora* and perhaps the parent calamites. Clearly, much more documentation of such relationships is required to further illuminate nuances of Calamitean paleoecology.

The sample intervals from the inner mire section (SH81) which are abundant (49.7-27.5 percent) in *Calamospora*, on the other, hand contain no megascopically discernible mudrock lithofacies. On the contrary, the two samples in which *Calamospora* are most abundant are composed entirely (SH81-6T) or predominantly (SH81-4T) of clarain (see Appendix D). The association between calamitean sphenopsids and humic, clarain-rich coal does not readily fit the interpreted ecological habitat deduced by Scott (1978, 1979). It is noteworthy, however, that these two intervals occur immediately above and below a decimetre-thick parting which marks the most southerly limit of splits derived from the basin-axial rivers.

The co-dominance of calamitean flora and arboreous lycopsids within the mire proper thus may reflect changing edaphic conditions related to periodic fluvial avulsion and abandonment. Why the dominant lycopsid tree flora declined through time, permitting the sphenopsids to expand, is problematic considering the obvious resilience of lycopsids in the piedmont zone, which was highly stressed by sheetflood. In the inner mire, the change to a sphenopsid-dominated flora was fostered by a decline in *Anabathra*, the parent tree of *L. orbicula*. The increase in sphenopsid miospores may alternatively record the establishment of the calamites within the diversified community without an actual decline in the number of lycopsid trees (DiMichele, personal communication, 1989).

4.13 THE NO. 3 SEAM: PROCESSES AND HISTORY OF MIRE DEVELOPMENT

Maceral-based interpretations (Calder *et al.*, 1991) indicate that the protected inner reaches of the ecosystem developed autogenically, resulting in modification to mire hydrology and chemistry. The suggested evolutionary trend to less groundwater influenced conditions is typical of many mires, both temperate and tropical. The piedmont (basin margin) and riverine (basin axis) zones developed under differing, hybrid processes of peat accumulation arising from interaction with allogenic processes of bordering deposystems (alluvial fan-piedmont and axial channel-belt). Subsequent incorporation of floral paleoecology and trends of ash, sulphur and lithotypes has permitted more accurate reconstruction of mire genesis (Figure 4.15), which differs somewhat from the maceral-based interpretation.

4.13.1 Inner Mire Processes

Interpretation of the evolution of the inner mire presents some intriguing dilemmas. The petrographic model of mire development of Calder *et al.* (1991) suggests a progressive evolution of lessening groundwater influence throughout the life of the mire. More precisely, the upward decrease in gelification, the major attribute measured by the groundwater influence index (GWI), indicates a steady decrease of pH level, O_2 and nutrient supply, inferred to accompany raising of the mire surface. The resulting evolutionary sequence of mire types was originally interpreted (Calder *et al., ibid.*) as swamp --> swamp forest --> fen, possibly culminating in oligotrophic bog and bog forest. The broader range of evidence considered in this paper, especially pertaining to floral ecology, suggests that the system remained under the influence of groundwater and probably never achieved a truly ombrotrophic status.

While there is compelling evidence of pervasive groundwater influence, the complete evolutionary sequence of mire types is less straightforward and may not have been one progressive continuum of less rheotrophic types as suggested by the steady decline of gelification. The upper third of the seam column (SH81) exhibits an increase in ash, sulphur and miospore diversity and a concomitant reversion to duller lithotypes, suggesting that the inner mire ultimately reverted to a more rheotrophic state in its latest stage of development (Figure 4.15). The resolution of these contradictory trends would require that the deflating phase of mire development remain acidic.

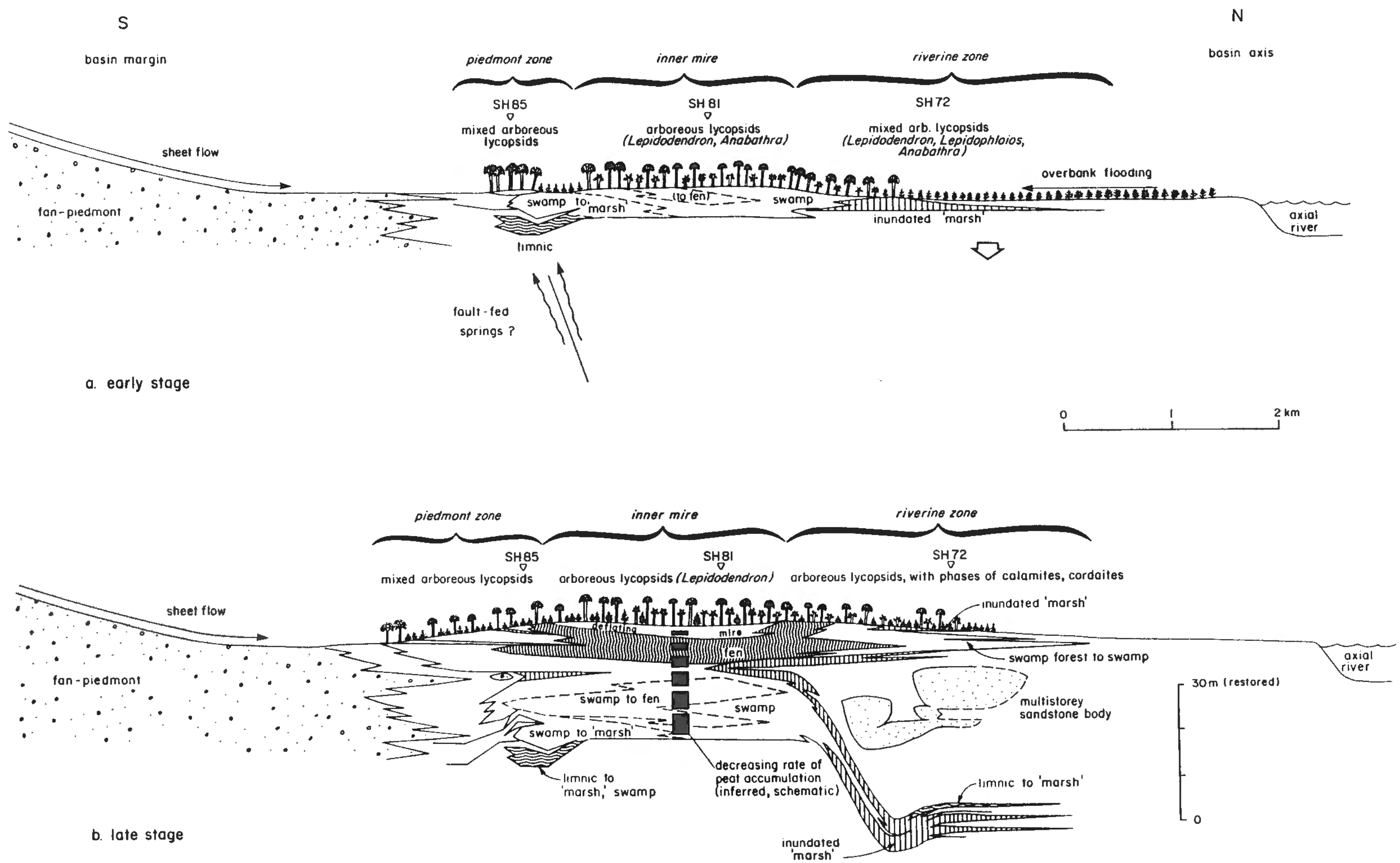


Figure 4.15 Evolutionary development of paleomire types for the ancestral peat-forming ecosystem of the No. 3 seam, Springhill coalfield: a) early rheotrophic (eutrophic) stage; b) late, post-mesotrophic (deflationary?) stage. This interpretation reflects reversal in trends of lithotype brightness, ash, sulphur and miospore species abundance at top of seam attributed to deflation rather than continued raising as suggested by maceral record alone. Restored thickness: peat (x7) mud (x2) sand (x1). (Topography exaggerated).

Upwardly decreasing rates of peat accumulation characterize the tropical forest mires of Southeast Asia as they grow ever less fertile (Anderson and Muller, 1975; Cameron *et al.*, 1989) and certain of these mires are undergoing rapid deflation (T.A. Moore, personal communication, 1990). Let us assume that the surface of the ancient Springhill systems may have actually deflated but alternatively may have accreted at a progressively slower rate in the inner mire region as a result of decreasing fertility. A decrease in rate of biomass accumulation would permit dilution of miospore species indigenous to the mire by externally derived spores (e.g. monosaccate pollen?, Figure 4.12). Similarly, the steady accumulation of wind and water borne ash would increase relative to diminishing peat accumulation. Nutrient stress in this mesotrophic system was probably much less pronounced than in modern oligotrophic examples, however. For such a modest change in nutrient supply to affect the rate of peat accumulation would therefore require that the dominant flora (*Lepidodendron* and *Anabathra*) be nutrient-sensitive in the extreme.

The probable late, regressive trend of inner mire evolution notwithstanding, the mire first records a mesotrophic tendency from swamp to fen. The succession of mire types within the ancient Springhill peatlands shows a similar tendency to the classic hydroseral succession of temperate zone mires that forms through the process of terrestrialization (Moore, 1987; *verlandung*: Weber, 1908) and to tropical zone mires of Indonesia, that form through "vertical paludification" (Cameron *et al.*, 1989; Esterle *et al.*, 1989). They differ in degree however, especially in that the No. 3 paleomire probably did not evolve to a solely rainfed hydrologic status.

According to the miospore record, floral succession within the inner mire appears to have been weakly developed, which is to be expected in a region with abundant water and nutrient supply (Grosse-Brauckmann, 1979; Teichmüller, 1989). Wholesale change in floral groups was rare: arboreous lycopsids in general dominated the mire throughout seam formation time. The apparent domination of mire vegetation by arboreous lycopsids, which required standing water cover for reproduction (Phillips, 1979), confirms that high groundwater levels prevailed throughout the life of the mire. The only major change in the dominant floral group (arboreous lycopsids to sphenopsids and cordaites) occurred periodically in the upper half of the seam, both in the inner mire (SH81) and riverine zones (SH72). Whether this represents a true autogenic succession is unknown.

Floral succession in the inner mire zone of the No. 3 seam is marked by species variation within a single genus or floral group, in particular the arboreous lycopsids, as opposed to variation at the scale of floral groups which suggests minimal ecological change. This succession comprised an initial *Anabathra/Lepidodendron hickii* community, with increasing domination of the mire flora by *Anabathra* before its sharp decline mid-seam. In the succeeding phase, *L. hickii* prevailed as the dominant plant, continuing through the final community, which was marked by the subordinate presence of *Lepidophloios harcourtii*. The domination of the mire flora by the two genera *Anabathra* and *Lepidodendron* is a reflection of the nutrient-rich, inundation-prone piedmont-riverine setting of the mire and supports current paleoecological interpretations of these plants. In the Westphalian B Low Barnsley seam of Yorkshire, Bartram (1987) interpreted *Lepidodendron* and *Anabathra (Paralycopodites)*-dominated peat as the early, colonizing and nutrient-rich rheotrophic phases (her 0 & 1) of mire development in a raised bog succession. *Anabathra* may be an indicator of the early rheotrophic, planar stage of peat formation in Westphalian mires elsewhere, given its similar occurrence within the relatively high ash, basal portions of the Westphalian D Herrin (DiMichele and Phillips, 1988) and Springfield (Willard, in press) coals of the Illinois Basin.

Densosporites, assumed to have been produced from herbaceous lycopsids, is uncommon not only within the No. 3 seam (Figures 4.12-4.14) but throughout the Cumberland Basin (G. Dolby, personal communication, 1989). These lycopsids, normally the climax vegetation of Westphalian B raised mires elsewhere (Smith, 1962; Bartram, 1987; Fulton, 1987; Eble and Grady, in press) and therefore assumed to be herbaceous by analogy to modern angiosperm-dominated mires, did not develop in this setting in the absence of pronounced, doming of the mire surface (cf. Figure 9 of McCabe, 1984, p. 28), linked either to prohibitive seasonality (Smith, 1962; Butterworth, 1966) or excessively rapid net subsidence (Fulton, 1987).

4.13.2 Piedmont Zone Processes

The piedmont zone of the mire initially developed under very high groundwater level, bordering limnic conditions. A weakly developed succession of mire types progressed only to the formation of a swamp, which prevailed for the life of the mire, albeit with episodes of inundated forest in response to piedmont incursions. The piedmont margin shows the weakest development of floral succession, perhaps due to its proximity to water- and nutrient- supplying fans (Grosse-Brauckmann, 1979); *Lepidodendron hickii* was the dominant miospore contributor. The progressive southerly onlap of brighter lithotypes (Figures 4.2, 4.3) indicates that the mire accreted in a progressive lateral manner toward the distal alluvial fans of the piedmont margin.

The process by which the mire developed in the piedmont zone is fundamental to an understanding of the origin of these and similar basin-margin/piedmont seams. The process of mire development at the piedmont margin, the site of fan-derived discharge and sheetflood, is analogous to the process of upslope paludification (cf. Verswampfung of Weber, 1908) described for the temperate mires of Polesie (Kulczynski, 1949). Here, the ecosystems develop progressively upslope by impeding the drainage of land of slightly higher topography. Peatlands in the Lake Agassiz region, Minnesota, developed in a similar manner (Heinselman, 1963; Tallis, 1983) termed lateral paludification by Cameron *et al.* (1989). The process was described by Heinselman as follows:

> "Upslope growth of the peatland seems to be achieved by damming up the incoming waters from mineral soils. This creates a wet area into which the swamp forest can advance and prepare the way for the bog flora. It is noteworthy that the initial invasion on mineral soils does occur in a wet area but the communities are still forests, not aquatics".

A similarly arboreous (lycopsid) flora characterized the piedmont margin of the No. 3 paleomire. Their dominance in a sheetflood-prone setting is in agreement with their interpreted tolerance to inundation so long as water cover did not become permanent (DiMichele and DeMaris, 1987), but, is apparently at odds with the interpreted intolerance of lycopsids to inundation by sediment (Gastaldo, 1986). In such a habitat prone to the stress of inundation, it is suggested that the plant group(s) with the most successful reproductive and/or growth strategy would have predominated.

The arboreous lycopsids had a heterosporous reproductive strategy (i.e. both microspores and megaspores), which utilized both wind and water transport for the dispersal of the large quantities of spores generated (Taylor, 1985; Phillips *et al.*, 1985; Collinson and Scott, 1987). They were thus well suited to become re-established in the aftermath of events such as fan-derived sheetfloods. An illuminating account of the growth and dynamics of a lycopsid forest is given in DiMichele and DeMaris (1987). The rapid growth of the trees, aided by their thick, supportive bark but little wood, and a single episode of mass reproduction (monocarpic) during late stage of growth, permitted them to establish on inundated sites, grow rapidly and once again reproduce. In addition, juvenile lycopsid trees (Figure 4.16) were unbranched (in *Lepidodendron aculeatum* until approximately 30 m high), permitting rapid growth without competition between the trees for sunlight (DiMichele and Demaris, *ibid*). The arboreous lycopsids were therefore particularly well adapted physiologically to growth in

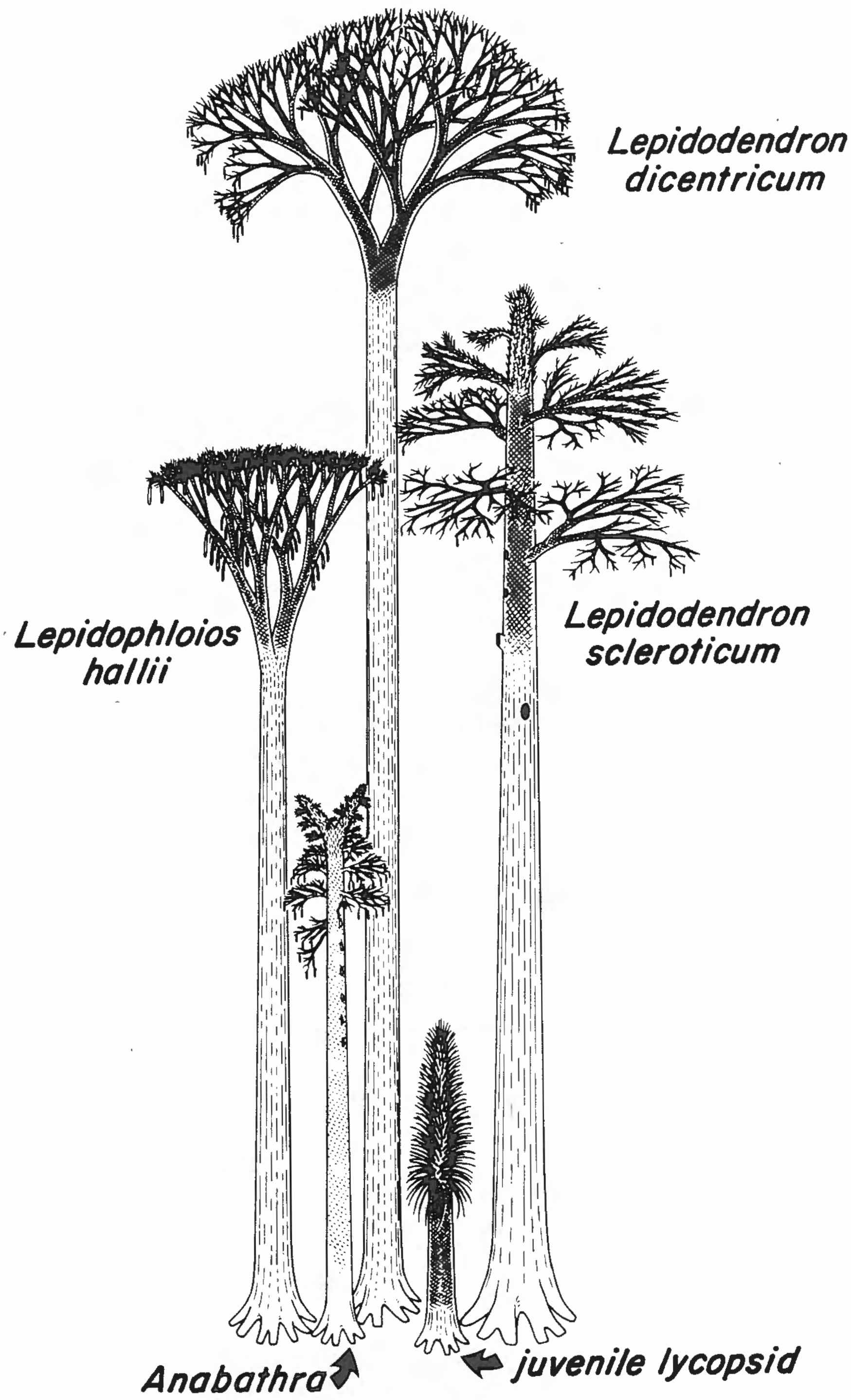


Figure 4.16 Reconstruction of arboreous lycopsids, after DiMichele and DeMaris (1987).

an intrinsically stressful environment such as the piedmont-riverine setting of the southern Cumberland Basin.

4.13.3 Riverine Zone Processes

Processes of peatland development at the riverine margin were profoundly influenced by channel-belt processes and by differential compaction. The conditions that prevailed in the riverine zone during accumulation of the lower leaf are of interest in that they were harbingers of later fluvial occupation. The peat accumulated under highly rheotrophic conditions from an arboreous lycopsid-dominated community (Figure 4.15), which in this area included *Lepidophloios hallii*, a plant that thrived in flooded regions of high abiotic stress (Phillips, 1979; DiMichele and Phillips, 1985; DiMichele *et al.*, 1985). Rotting indicated by the presence of fungal mycelia (Dolby, 1988a), unique to the lower leaf (bench) of the riverine zone may be an indicator of such stress, but its ecologic significance is not entirely clear. The mire was inundated and was the site of mud deposition, represented by 2-5 m-thick mudrock, prior to the site being occupied by the sand-laden axial river, suggesting that an underlying process controlled both mire development and channel-belt migration in this zone. Contributing to this may have been the differential compaction of multistorey sandstone bodies and mudrocks. Evidence in support of this hypothesis is found in the sharp contrast in maceral composition of the upper and lower benches, the upper bench indicating peat accumulation under significantly less influence of groundwater (swamp and fen), inferring a higher topography for the riverine zone mire above the entombed multistorey sandstone body. Another potential control is syndepositional faulting, suggested by the presence in abandoned mine workings of a fault paralleling the southerly margin of fluvial occupation not only in the No. 3 seam, but in the underlying No. 1 seam as well.

The ultimate demise of the ecosystem is problematic and a convincing allogenic control has yet to be identified. The cessation of peat accumulation in many marine-influenced settings can be convincingly attributed to combined climate and sea level change, but for intermontane basins with no apparent marine influence the mechanism is less clear. Steadily decreasing rate of biomass production, hence peat accumulation, may have accompanied inferred decrease in mire fertility (nutrient supply) and may have reached a critical value relative to net subsidence of the basin. As a result, the mire may have become progressively susceptible to allogenic change as it fell into disequilibrium with surrounding environmental conditions such as basin subsidence or climate change.

4.14 CONCLUSIONS

Evidence of changing groundwater influence is the fundamental parameter for interpreting the genesis of modern and ancient peatlands. Changes in groundwater influence in the No. 3 seam are reflected in the distribution of vitrinite maceral types and especially the gelification of tissues; the ecology and succession of mire floras; the distribution of detrital mineral matter and vertical trends of lithotypes. Access to the record stored in the vitrinite macerals requires resolution of the progressive gelification of tissues. At the very least, this entails recognition of the macerals and types telinite, telocollinite, gelocollinite and corpocollinite, although even these terms may be insufficient to fully describe the spectrum of tissue preservation and the products of various pathways of degradation. Etching by oxidative agents should provide important insight into the degree to which cryptic tissue structure has suffered the effects of biochemical gelification (e.g. Warwick and Stanton, 1988; Moore *et al.*, 1990) Reconstructive diagrams of mire type and evolution based on maceral data are useful tools in paleo-environmental analysis, but it is important to recognize the fact that they are inherently limited in their depiction of the complex evolution of peat-forming ecosystems.

The paleoecology of the mire flora and the record of floral change are fundamentally important to the interpretation of ancient peat-forming ecosystems. Furthermore, paleobotanical data is required to verify processes of degradation which can be profoundly affected by tissue type.

The ancestral paleomire of the No. 3 seam developed as a predominantly rheotrophic system, although there is evidence that it tended toward lesser groundwater influence as do many peat-forming ecosystems. The predominance of arboreous lycopsids, particularly *Lepidodendron* and *Anabathra*, confirms the pervasive influence of nutrient-rich groundwater. Domination of the mire flora by these arboreous lycopsids and the minimal development of floral succession is atypical of many Westphalian mires, where such nutrient-rich, arboreous lycopsid peat forms only a portion of the overall mire succession, usually in early stages of development. Interpretation of mire genesis confirms the primary, fundamental control that groundwater exerted on the development of peat (and ultimately coal) -forming ecosystems in the Springhill coalfield.

Many studies of ancient peat formation tend to focus on "inner mire" zones, rather than on subeconomic marginal zones. Processes of mire development may differ in the inner and peripheral regions. The margins provide insight into the interaction of the peatlands with surrounding deposystems, thus providing a clearer understanding of the ecosystem as a whole, and controls on its development. The inner mire of the No. 3 seam evolved through partial terrestrialization (cf. vertical paludification of Cameron *et al.*, 1989), and may have ultimately reverted (deflated) to a more groundwater-influenced condition. The feedback mechanism of lateral paludification at the piedmont margin was aided by the dispersed nature and low competence of fan-derived groundwaters and especially by groundwater seepage from distal fans, and by the ecological adaptation and growth strategy of arboreous lycopsids. Peat formation at the riverine margin was profoundly influenced by allogenic processes including differential compaction around multistorey sandstone bodies, and by channel-belt processes, which in turn may have been locally influenced by contemporaneous faulting.

The ultimate demise of the peat-forming ecosystems in this intermontane setting remains one of the most perplexing aspects of their development, but the rapidity of basin subsidence must have had a strong influence on the ecosystems that developed within the Cumberland basin.

CHAPTER V: DISCUSSION OF CONTROLS ON ANCIENT PEAT ACCUMULATION IN THE SPRINGHILL COALFIELD

Numerous, essentially unlinked, observations have been made thus far in the course of this study. These include: the stratigraphic position of seams within the basin-fill record (the so-called "coal window"); the areal distribution of seams and in particular the inferred reliance of the peat-forming ecosystems on groundwater discharge from alluvial fans; coincidental position of coal seams, multistorey sandstone bodies and faults; intimate association of coal and multistorey sandstone within the so-called cyclothem; evidence of interaction of mires with neighbouring deposystems (piedmont alluvial fans and basin-axis rivers); the specialized paleoecology of peat-forming floras; geometric, lithologic, chemical and floral zonation of seams, all with their attendant interpretations.

The task at hand is to link the process or processes responsible for each of these and numerous other observations and, subsequently, to weigh them in a hierarchical order of controls on ancient mire development. The majority of observations were interpreted in earlier chapters, however it is not possible to come to an objective conclusion on controls until all available evidence can be considered. Importantly, if ecosystems today serve as an example, nature will prove to have been in delicate balance, and few, if any, of the identified controls will have acted independently of all others.

5.1 TEMPORAL RANGE OF CONTROLS

One method of inferring controls on ancient peat formation with an eye to relative weighting is to consider relative time frames of observed features (e.g. cycles and events) and to compare these with empirical time frames of climatic, tectonic and other processes. Thus, an objective methodology can be applied in the hope of achieving some semblance of order from the chaos of processes and controls.

5.1.1 Constraints in the Interpretation of Time

The interpretation of such controls is severely limited by our ability to deduce time in the ancient record. There are no absolute radiometric dates from the Carboniferous basins of the Maritimes, therefore, one is forced to rely upon biostratigraphic correlation with basins from which such data are available, chiefly in NW Europe. The radiometric dates currently considered to be best constrained (Leeder and McMahon, 1988; Klein, 1990) are those derived from sanidine within tonstein beds in NW Europe (Lippolt and Hess, 1985; Hess and Lippolt, 1986), which give the duration of the Westphalian stage at just 10 Ma (Figure 1.2), in comparison with the previous and widely accepted time scale of Harland *et al.* (1982) which gave the Westphalian a much greater duration of 19 Ma. Depending on the time scale utilized, significantly different interpretations can be drawn with respect to cyclicity (cf. Heckel, 1986 and Klein, 1990), and caution is required in interpolating time spans of cycles in the order of 10^4 - 10^5 years when the constraining radiometric dates have a margin of error in the 4 x 10^6 range (Klein, 1990).

Nonetheless, time is the key to the ultimate understanding of processes and cyclicity. The timing of events and cycles relative to one another is less vulnerable to the above-mentioned pitfalls than is the acquisition of absolute dates, however other limitations to our understanding of time loom large. The major constraint is the measure of time represented by a bituminous coal seam. Such an interpretation is based mainly on two variables: i) the rate of peat accumulation; and ii) the compaction ratio of the peat to bituminous coal.

The rate of Carboniferous peat accumulation is unknown and undoubtedly varied, as today, between mire types and geographic locations. One is forced to accept the Law of Uniformitarianism and deduce rates based on modern examples, which range from 0.3-1.0 mm yr^{-1} for the warm temperate Okefenokee swamp (Spackman *et al.*, 1976) and 0.8 mm yr^{-1} for the warm temperate Everglades (Spackman *et al.*, 1969) to 2.8 mm yr^{-1} for the equatorial forest mires of Southeast Asia (Anderson, 1983). Whereas the arboreous lycopsids, chief contributors to the peats of the Springhill coalfield, were of efficient construction suggestive of very rapid growth (DiMichele and DeMaris, 1987; Stewart, 1983), the higher rate of accumulation for tropical peats of Indonesia is considered to be a closer analogy than the lower rates for temperate mires of southeast USA. It should be noted, however, that Indonesian peats, which are composed predominantly of wood, are not precise analogues for Carboniferous peats, because Carboniferous floras contributed bark but little wood to the peat (DiMichele, written communication, 1990). Other factors, such as the evolution of detritivores (Raymond, 1989) and lignin-decomposing fungi (Robinson, 1990), would also have influenced this rate.

Estimated peat: coal compaction ratios vary widely from 1.4:1 to 30:1 (Ryer and Langer, 1980). The potential impact of original peat biomass on the range of compaction ratios was well illustrated by Winston (1986) who determined differential compaction of tissues: cordaitean wood at 3.3:1, lycopsid periderm at 7:1 and stigmarian rootlet at 30:1. Moore and Hilbert (in press) on the other hand, found little change in compaction of tissues between a Holocene peat and Miocene lignite from Indonesia. They inferred that compaction arose chiefly from loss of pore space and water. The inherent problems of using a standard compaction ratio are reviewed by Collinson and Scott (1987). Perhaps the most commonly utilized estimate is the 7:1 ratio suggested by Teichmüller (1982), which is also the average of published ratios compiled by Ryer and Langer (1980).

Even when constrained by compaction ratios ranging from 4:1 to 10:1 and accumulation ratios of 0.8 to 2.8 mm yr^{-1} (Table 5.1), calculations of time represented by one metre of bituminous coal range from 1700 to 12,500 years. An important point to be considered in choosing a decompaction ratio for coal, so as to deduce time, is that the so-called peat accumulation rates for modern tropical mires are the time represented by 1 mm of variously compacted peat averaged over a given thickness interval constrained by radiometric dating. The time represented by a given peat interval is not equivalent to the actual rate of biomass accumulation at the surface, because the peat will have already experienced some degree of compaction. The basal layers of a peat deposit subjected to radiometric dating will be especially compacted. Therefore, it may be advisable to utilize a somewhat lower ratio of compaction, when coupling this with radiometrically based ages ("accumulation rates").

Compaction ratios from well compacted and coal ball-bearing peat to coal given by Ryer and Langer (1980) range from 2.2:1 (Stutzer, 1940) to 10:1 (Lewis, 1934; Bloom, 1964) and average approximately 5:1. Using this ratio and the average accumulation rates cited for tropical forest mires (2 mm yr^{-1}), the time represented by one metre of bituminous coal might therefore be as little as 2500 years (Table 5.1), as opposed to the 6000 year estimate given in Stach *et al.* (1982).

5.1.1.1 Time represented by the basin-fill sequences and cyclothems

The relative time span of sequences within the basin-fill record of the Springhill coalfield is depicted in Figure 5.1. The basin-fill time frame was deduced by palynological correlation of the basin-fill sequence with the European stages (Westphalian) (Dolby, 1991) and thence by utilizing the Ar^{39}/Ar^{40} based time scale of Lippolt and Hess (1985). The basin-fill sequence (Figure 3.14), comprising the basal conglomerates (Polly Brook Formation) and coal-bearing strata (Springhill Mines Formation) of the Cumberland Group, of late Westphalian A to mid-Westphalian B age, may represent a time period of 2-3 Ma. The entire Cumberland Group, as redefined (Ryan *et al.* in press), represents approximately 4 Ma. The temporal range of cyclothems, and smaller cycles and events, was deduced by two methods: 1) from time assumed for the formation of coal, as discussed previously, and 2) by dividing the number of cyclothems into the basin-fill. The two methods, fraught with

Peat Accumulation Rate (mm, yr^{-1})	Mire Type	Assumed Compaction Ratio	Time (Years per m bituminous coal)
		5:1	5,000 - 13,330
0.3 - 1.0	swamp forest,	7:1	7,000 - 23,330
	Okefenokee, Georgia[a]	10:1	10,000 - 33,330
		20:1	20,000 - 66,670
		5:1	6,250
0.8	swamp forest,	7:1	8,750
	Evergrades, Florida[b]	10:1	12,500
		20:1	25,000
		5:1	1,790
2.8	bog forest,	7:1	2,500
	Sarawak[c]	10:1	3,570
		20:1	7,140
		5:1	2,630 - 4,170
1.2 - 1.9	bog forest,	7:1	3,680 - 5,830
	Sarawak[d]	10:1	5,260 - 8,330
		20:1	8,420 - 13,330
		5:1	2,500
2.0	bog forest,	7:1	3,500
	Sarawak (avg.)	10:1	5,000
		20:1	10,000

[a]Spackman *et al.*, 1976
[b]Spackman *et al.*, 1969
[c]Anderson, 1983
[d]Cameron *et al.*, 1989

Table 5.1 Peat accumulation rates from various subtropical-tropical mires, and possible time represented by 1 metre of bituminous coal.

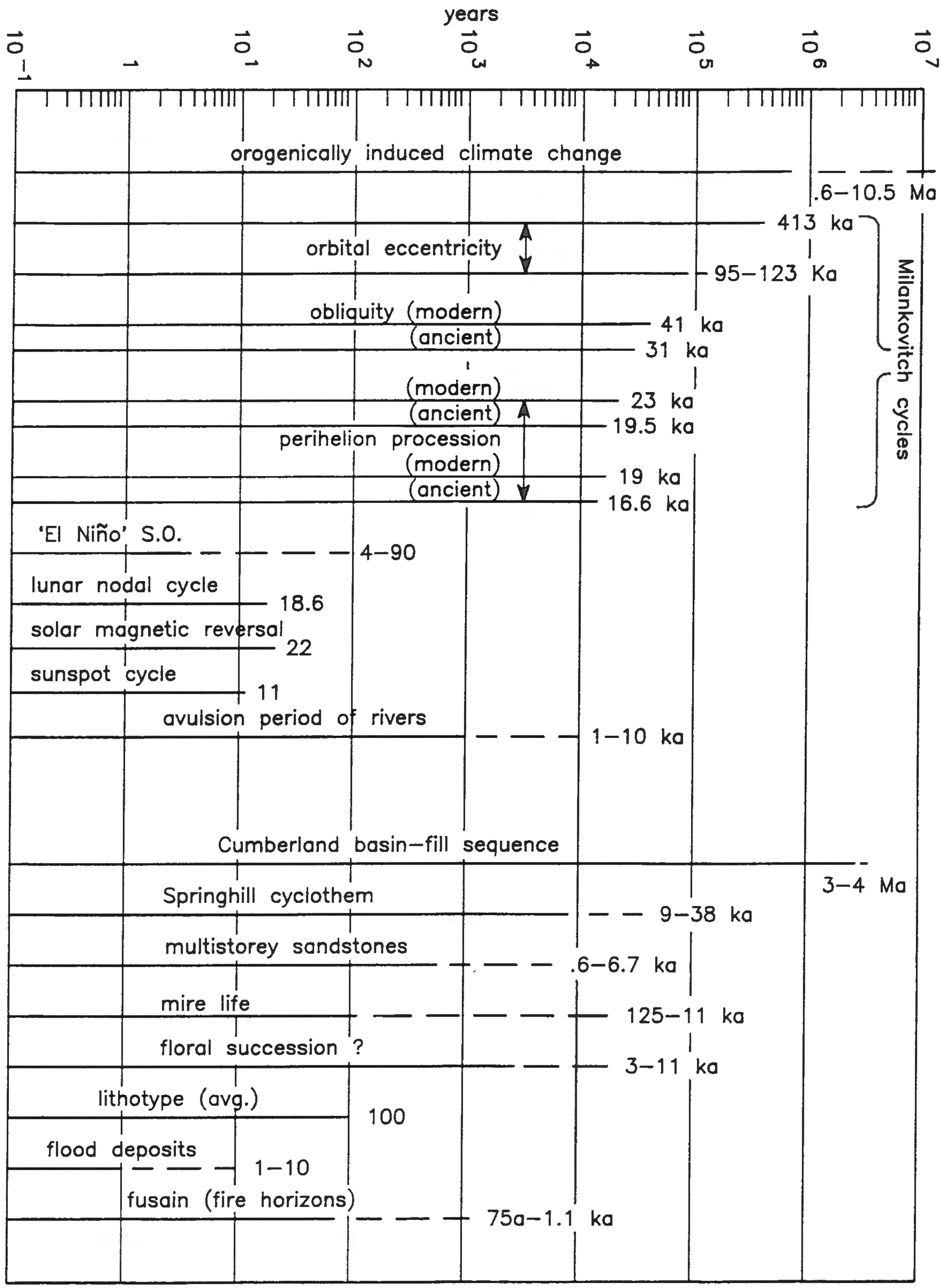


Figure 5.1 Time frames of natural cycles and processes relative to those deduced for the coal-bearing strata of the Springhill coalfield.

assumption, show general agreement when aggregate time represented by all potential cyclothems is enumerated.

Time represented by coal seams in the Springhill coalfield, which range from 5 cm to 4.3 m thick, is deduced to have ranged from 125 to 10,750 years. The relative time frame for deposition of a multistorey sandstone and overlying and underlying mudrocks was deduced by measuring the total uninterrupted seam thickness of the No. 3 seam and comparing this with the total of two leaves of the seam between which is entombed a 22 m-thick parting including a 13 m-thick multistorey sandstone body (Figure 4.2). The sum of the leaves (1.55 + 1.3 m = 2.85 m) subtracted from the thickness of peripheral, uninterrupted coal (4.4 m) yields a difference of 1.55 m, which may represent 3875 yrs. [Such a calculation necessarily assumes that the history of peat accumulation was constant across the mire, however radiometric dates from various profiles across a modern forest bog in the Baram River area of Sarawak indicate that the onset of peat accumulation differed as much as 1470 yrs. over a distance of 17 km (Cameron *et al.*, 1989). Furthermore, a discontinuity of 1155 yrs. was defined within the peat arising from infilling of an abandoned mire channel]. By this method, the total time represented by a coal-mudrock-multistorey sandstone-mudrock cyclothem, with reservations, may have ranged from 9-15 ka for the major seams of the Springhill coalfield.

An alternate method of deducing the period of time represented by a cyclothem (cf. Chesnut, 1989; Gibling and Bird, in press) is to divide the number of cyclothems into the biostratigraphically constrained basin-fill sequence for which an absolute age has been inferred through correlation with the European stages. (Unfortunately, no absolute radiometric ages have been determined from the Cumberland Basin fill, or from the coal measures elsewhere in the Maritimes Basin.)

The Westphalian A-B boundary in the Springhill coalfield (see section 3.1.2, Chapter III) has been placed between the No. 2 and the No. 7 seam , most probably within an interval between the No. 2 seam and 170 m below (Dolby, 1991). An upper boundary of the basin-fill at Springhill is more problematic. The coal-bearing strata at Springhill are believed to be no younger than early to mid Westphalian B (Dolby, 1991). The drilling of the deep corehole SA.88.2 in the mid-region of the Athol Syncline in 1988-89 (Calder and Gillis, 1989) provided important coverage of strata that overlie the Springhill sequence, including the youngest strata dated within the Athol Syncline (Dolby, 1991). Importantly, the South Athol corehole intersected a sequence of coal seams at depth that have been correlated with the No. 2 to Rodney seam interval (Dolby, 1991; Calder and Bromley, in press), thus providing a stratigraphic continuum with the Springhill coalfield sequence. The strata intersected in the South Athol hole are believed to extend to very latest Westphalian B (Dolby, 1991) thus the composite Springhill-South Athol basin-fill sequence spans virtually the entire Westphalian B. The thickness of the composite Westphalian B basin-fill sequence is 1150-1500 m, given the uncertainties discussed above in the precise correlation between South Athol and Springhill and the position of the Westphalian A-B boundary. The Westphalian B/C boundary in Europe has been dated at 311 Ma (Lippolt and Hess, 1985), but the A/B boundary has not been dated. Assuming that the Westphalian A and B were of equal duration (e.g. Leeder and McMahon, 1988), the Westphalian B may have lasted 2 Ma, but this time frame must be treated with considerable scepticism until such time as it becomes more rigidly constrained.

Assuming a constant rate of sedimentation throughout the Westphalian B, each cyclothem in the Springhill coalfield (averaging 24 m) may represent a time period of 32,000-38,400 years. These time frames are double to quadruple the 9-15 ka period deduced through inferred peat accumulation rates and compaction ratios, although the results of the two methods at least fall within the same order of magnitude (10^4 yrs.). It appears unlikely therefore, that the time frame represented by a Springhill cyclothem is similar to the 200 to 400 ka periods (10^5 yrs.) suggested for cyclothems elsewhere (e.g Heckel, 1986; Chesnut, 1989; Chesnut and Cobb, 1989; Klein, 1990; Gibling and Bird, in press).

5.1.1.2 Time represented by small-scale phenomena

The subtle nature and possible regression of floral succession within the No. 3 seam makes it difficult to assess the time span of succession. Although it may not have spanned the entire life of the mire as suggested by McCabe (1984), floral succession probably occurred progressively throughout much of the mire's development. Peat accumulation in Westphalian mires of Britain appears to have continued for some time after the climax *Densosporites* phase (Smith, 1962) developed. The climax vegetation in these seams was reached within 1 to 1.8 m of coal, possibly representing 2.5-4.5 ka.

The average time required for the development of a macroscopic lithotype (4.8 cm avg.) in the No. 3 seam is deduced to have been in the order of one to two centuries. Clearly, certain lithotypes would have formed in much less time, an example being discrete vitrain derived from bark of a single fallen lycopsid tree. Discrete fusain layers within the inner mire zone of the No. 3 seam formed at intervals ranging from 75 to 1100 yrs., averaging 400 yrs.

5.2 EARLY WESTPHALIAN PALEOCLIMATE

The existence of the extensive Late Carboniferous peatlands of Euramerica has been attributed to a favourable inter-continental configuration that is thought to recur in *ca.* 400 Ma periods, resulting in alternating "greenhouse" and "icehouse" climates (Fischer, 1984, 1986; Veevers, 1990).

Paleoclimate reconstruction for the Euramerican province during the Late Carboniferous (Figure 5.2) has focused in the main upon the Appalachian coal basins of USA (Cecil *et al.*, 1985; Cecil, 1990; Winston and Stanton, 1989) with no published attempt at a comprehensive paleoclimatic interpretation for the Maritimes Basin. An exception has been the work of van de Poll (1978), who interpreted an arid climate during the Early Carboniferous with a return to more humid conditions during the Late Mississippian-Early Pennsylvanian, with seasonal changes in precipitation and warm subtropical temperatures. A comprehensive paleoclimatic study is beyond the scope of this dissertation. Climate, however, is a fundamental allogenic control on peatland development (Gore, 1983; Phillips and Peppers, 1984; Cecil *et al.*, 1985). It is necessary therefore to re-appraise the Westphalian paleoclimate, in particular in light of evidence secured during the course of this research.

Studies of Euramerican coal province climate during the Westphalian (Figure 5.2) have followed three main lines of approach: i) floral change (Phillips and Peppers, 1984; Winston and Stanton, 1989); ii) lithology and inferred mire topography (domed or planar; Cecil *et al.*, 1985); and iii) continental and oceanic reconstruction and atmospheric circulation (Rowley *et al.*, 1985).

On the basis of ecological change inferred from the record of peatland and other ecosystem floras, Phillips and Peppers (1984) deduced five major climatic periods for the Euramerican coal belt during the Late Carboniferous: i) wet during the Westphalian A; ii) dry during Westphalian B-C; iii) wettest during Westphalian D; iv) driest during Stephanian, v) becoming wetter in late Stephanian time. Further studies by Phillips *et al.* (1985) led to the interpretation of a seasonal paleoclimate for the Westphalian B that was relatively drier than the Westphalian A.

In contrast, the Westphalian B paleoclimate of the central Appalachian Basin has been interpreted as everwet tropical (Cecil *et al.*, 1985). In a subsequent publication, (Cecil, 1990) long-term climatic trends were interpreted as: i) semi-arid from Visean through Namurian A; ii) rainy from Namurian B through Westphalian B; iii) increasingly dry during Westphalian C-D becoming iv) semi-

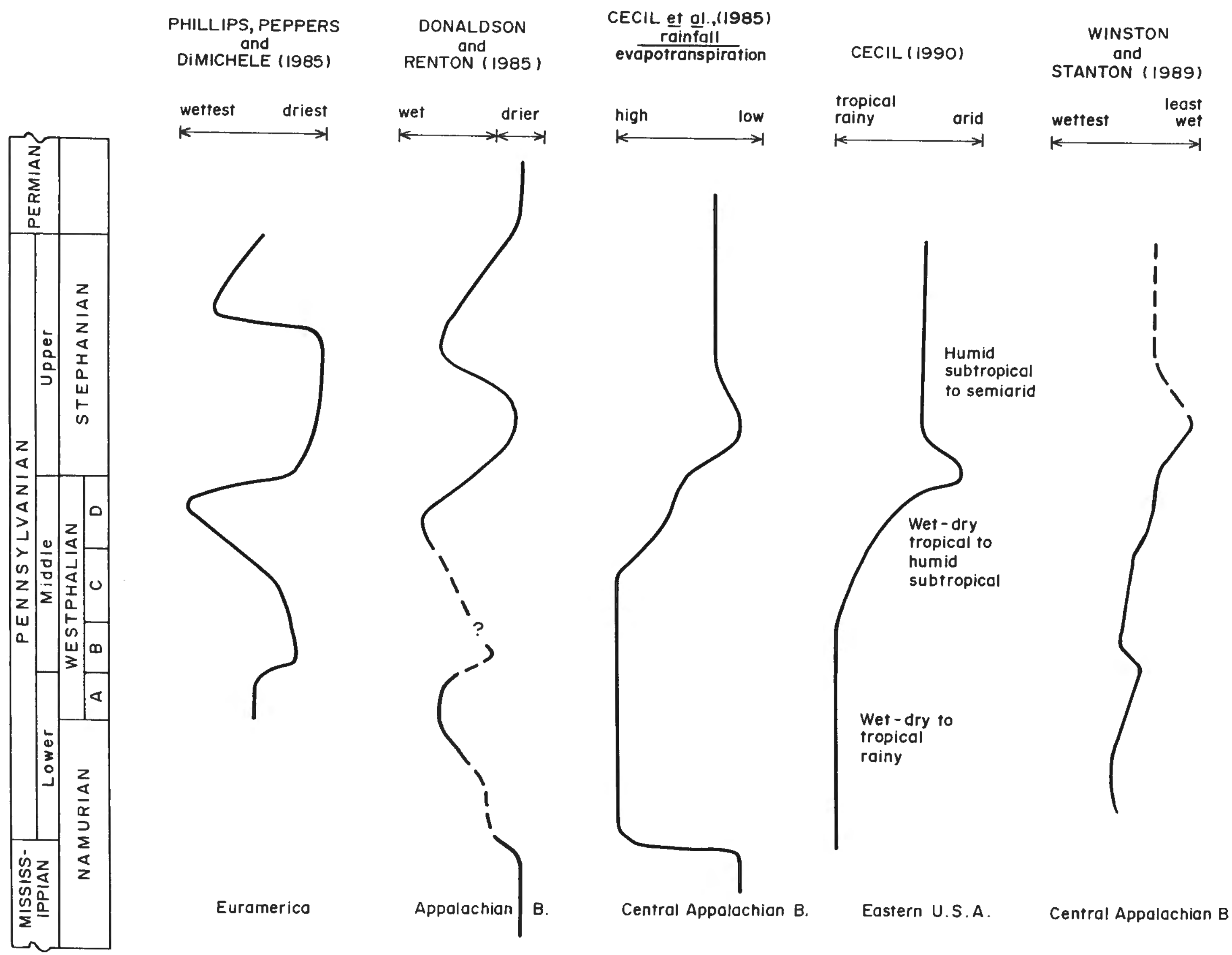


Figure 5.2 Carboniferous paleoclimate interpretations for various areas and basins of Euramerica.

arid in Stephanian time. Cecil (1990) allowed that the Namurian B-Westphalian B paleoclimate may have shifted periodically from (tropical) rainy to long wet/short dry seasonal. The formation of coals, presumably derived from domed peatlands, was linked to rainy cycles with enhanced siliciclastic deposition during seasonal periods. Coarsening upward basin-fill sequences and the presence of splint coals interpreted as domed peat deposits subjected to periodic aerobic degradation (Eble and Grady, 1988) were cited as evidence for increasingly frequent seasonal periods in the central Appalachian Basin during the Westphalian C-D as Euramerica drifted out of the intertropical convergence zone into monsoonal latitudes.

The paleoclimate curve deduced by Winston and Stanton (1989), based largely on evidence of floral change, incorporates findings of other researchers across Euramerica. Their curve is similar to that of Cecil *et al.* (1985) and Cecil (1990), but suggests a gradual drying trend throughout the Westphalian and Pennsylvanian, with a relatively drier interval during the Westphalian B (cf. Smith, 1964; Phillips *et al.*, 1985; Donaldson and Renton, 1985).

The paleoclimate of the Cumberland Basin need not have been similar to that of the central Appalachian Basin, due to a different position relative to the rising Appalachian mountains and to the ocean. A low pressure cell that would have developed over the Appalachians (Rowley *et al.*, 1985) may have drawn moisture-laden easterly winds across the Maritimes Basin. The direction and relative moisture of atmospheric circulation is a function of a complex interplay of pressure cells of which the Appalachian cell would have been but one. If the low was of sufficient intensity to draw airflow over the Maritimes Basin, the prevailing flow may have been northeasterly given clockwise circulation in the southern hemisphere. Were the Cobequid highlands oriented east-west as today, the piedmont mires would have been in a position to benefit from a rainshadow effect. Assessment of atmospheric circulation for the Maritimes Basin, however, is presently constrained by uncertainties in paleogeographic reconstruction, especially with respect to western Europe (especially the Iberian peninsula and Great Britain).

During the Late Carboniferous, the continent of Gondwana lay adjacent to the Maritimes Basin and experienced several glaciations in the polar region (Crowell, 1978). Although the Cumberland Basin lay in an equatorial setting, fractures on quartz grains from the Parrsboro Formation of early Westphalian age have been interpreted to be of glacial origin (D'Orsay and van de Poll, 1985). Early forms of bisaccate pollen from the Westphalian B strata of the Springhill coalfield and especially from coeval coal-bearing strata of the Debert-Kemptown coalfield on the opposite (south) side of the Cobequid highlands resemble cold weather Gondwana flora (G. Dolby, personal communication). Such evidence is suggestive of a local cold Alpine climate presumably associated with the Cobequid highlands, although this mountainous region has not generally been thought of as being of such an elevation. Equatorial peat-forming rainforests and highland glaciers coexist within 150 km of one another in Papua New Guinea.

Ross and Ross (1988) attributed extensive preservation of coal in the late Namurian through Westphalian (Middle Carboniferous) to low bacterial action and high rates of carbon production in a cool climate. In support of cool temperature they cited a substantial drop in shallow shelf marine water temperature (ca 18°C to 10°C) recorded in Ca/Mg ratios of invertebrate fauna (Yasamonov, 1981) and contended that the extinction of shallow marine calcareous invertebrates at the end of the Westphalian, beginning of Stephanian, may have been due to rapid climatic cooling.

Ross and Ross (*ibid.*) also contended that paralic coals favoured low average groundwater levels which resulted from the extended Gondwana glaciation (Crowell, 1978), and inferred that sea level rise associated with global warming favoured preferential preservation of limnic (intermontane) coals. If this hypothesis is true, the intermontane, "limnic" coals of Port Hood, St. Rose, Joggins, Springhill and Stellarton, would reflect a period of global warming from mid Westphalian A through C, with a cooling trend in Westphalian C through Stephanian time wherein the paralic coals of the Sydney Basin (Hacquebard and Donaldson, 1969) were formed. It would seem to be implied, but not stated, in the model of Ross and Ross that limnic coals would co-exist with paralic coals during periods of low sea level. The absence of intermontane (limnic, *sensu latu*) coals during the later Westphalian, however, suggests that regional tectonism was an important factor in the formation of these two main coal settings in the Maritimes Basin.

Many researchers of Euroamerican paleoclimate (Smith, 1962; Phillips *et al.*, 1985; Donaldson and Renton, 1985; Winston and Stanton, 1989) have inferred that the Westphalian B experienced a change to relatively drier conditions. The poor representation of Westphalian B strata across the Maritimes Basin renders a regional assessment of paleoclimate for this time difficult. It is known, however, that the Westphalian B was a tectonically active period, marked by active thrusting and alluvial fan building in southeastern New Brunswick (Plint and van de Poll, 1982, 1984; Nance, 1987), by a great accumulation of basin-fill in the Cumberland Basin, and largely by regional erosion elsewhere. It is possible, therefore, that local orographic effects (rainshadows) played a role in determining the local paleoclimate of the Cumberland Basin and similarly may have affected other regions of the Maritimes Basin at various stages throughout the Westphalian.

The Cumberland Basin may have experienced a similar overall drying trend throughout the Westphalian C-D and on into the Stephanian as that inferred for the Appalachian Basin (Cecil *et al.*, 1985; Winston and Stanton, 1989; Cecil, 1990). Indeed, much of the southern Maritimes Basin may have experienced such a trend, as evidenced by the widespread red beds across much of New Brunswick, northern Nova Scotia and Prince Edward Island. During the Westphalian D, however, the Sydney and Gulf of St. Lawrence Basins were marked by widespread peat accumulation (Hacquebard, 1986) suggesting different conditions there. A closer proximity to the sea (Gibling and Bird, in press) may have resulted in higher atmospheric moisture and elevated water tables in the eastern Maritimes Basin.

5.2.1 Rheotrophic and Ombrotrophic Mires as Climatic Indicators

Ombrotrophic, raised mires receive water solely from rainfall and are indicators of relatively evenly distributed annual precipitation and low evapotranspiration. Rheotrophic, planar to weakly domed mires, on the other hand, are more likely to occur in areas of seasonal precipitation where groundwater recharge forms an important dry season supplement to water influx (Gore, 1983; Moore, 1987).

Planar and domed peat-derived coals have been used as major criteria in distinguishing seasonally dry from everwet paleoclimates (Cecil *et al.*, 1985; Cecil, 1990). Their use, however, is contingent upon our ability to identify correctly these basic mire types in coal beds. Divergent methodology and ensuing interpretation in the recognition of ancient rain-fed vs. groundwater-fed mires (cf. Eble and Grady, 1988; Grady and Eble, 1990; Calder *et al.*, 1991; Calder, 1989; in press) jeopardize their usefulness as paleoclimatic indicators. For example, the seemingly paradoxical interpretation of an aerobically degraded raised mire (Eble and Grady, 1988) led Cecil (1990) to cite a domed (hence, rainfed) mire as an indicator of a seasonally *dry* paleoclimate in contradiction to the promising hypothesis of Cecil *et al.*, (1985) that domed, rainfed peat/coal beds are representative of low evapotranspiration: precipitation, or *humid* conditions. The major hindrance to the interpretation of raised and planar mires is in the limitation of coal petrography to recognize evidence of trophic status, imposed by restrictive maceral terminology and by limited knowledge of maceral genesis. This is particularly so for the vitrinite, but also the inertinite maceral group (Calder *et al.* 1991), and is compounded by petrographic terms used in studies of modern analogues which are not readily transferable to conventional maceral terminology (Esterle *et al.*, 1989).

5.2.2 The Densoporites Question

Smith (1962, 1968) identified four phases in the development of the Westphalian A and B coals of Yorkshire: Lycospore, Transition, Densospore and Incursion. The Lycospore through Transition to Densospore phases were postulated by Smith to record the evolution of a raised mire, and hence rheotrophic to ombrotrophic conditions. The Incursion phase was considered to arise from a flood event which might occur at any time during this development. The Densospore phase is characterized by the dominance of *Densosporites spp.*, which is rare not only in the Number 3 seam (Dolby, 1988a) but within the Cumberland Basin generally (Hacquebard and Donaldson, 1964; Dolby 1991). Smith (1962) presented two hypotheses for the development of the Densospore phase:

1) a raising of the mire surface in a constantly humid climate. The Densospore phase reflects a climax vegetation of herbaceous lycopsids which were specialized to grow in relatively highly acidic and low-nutrient (oligotrophic) conditions; and

2) climatic change: increase in precipitation, humidity and possibly temperature; an analogue is the blanket bog.

Both hypotheses require relatively high precipitation: evaporation ratio and low seasonality (Gore, 1983; Frenzl, 1984; Thompson and Hamilton, 1983; Ingram, 1983).

Smith (1962) and Butterworth (1966) suggested that the absence of the Densospore phase in the Westphalian C coals of Britain reflects an unsuitable (i.e. relatively drier) paleoclimate. This interpretation was questioned by Fulton (1987), citing the conflicting paleoclimate interpretation of Phillips and Peppers (1984). A fundamental point overlooked by Fulton was that the model of Phillips and Peppers, while extrapolated to include the Euramerican coal province, was focused on the Appalachian Basin. As discussed earlier, the paleoclimate for areas east and west of the rising equatorial Alleghenian mountains in all likelihood would have differed.

The occurrence of *Densosporites sp.* in Westphalian D coals of Pennsylvania has been linked to increased salinity due to marine transgression (Habib and Groth, 1967). Fulton (1987) rejected this theory on the grounds that *Densosporites sp.* occurs within the non-marine associated "Thick

Coal" seam of the Warwickshire coalfield, U.K. The possibility that a broader interpretation of pH dependence (i.e. alkaline conditions) could be invoked to explain the occurrence of the genus must be similarly dismissed due to its rarity within the Number 3 seam.

Does the paucity of *Densosporites sp.* mean, then, that there was insufficiently abundant and steady precipitation to permit development of raised mires in the Cumberland Basin during the Westphalian B?. Not necessarily, according to recent studies of miospore successions from the Westphalian B of the Warwickshire coalfield (Fulton, 1987). Fulton found *Densosporites sp.* to represent a raised mire climax vegetation whose formation was dependent upon relatively low subsidence rates and long-term stability of the peat accumulation/subsidence ratio. The resulting organic facies produced under these conditions was termed a "long-residence histosol". One inference based upon such a model is that basin subsidence relative to peat accumulation was too great for the development of the *Densosporites sp.*-producing community in the Cumberland Basin.

5.3 PALEOCLIMATIC EVIDENCE FROM THE SPRINGHILL COALFIELD

Evidence in support of an inequable, perhaps seasonal, climate for the Springhill coalfield (Westphalian B) includes: a) the rheotrophic to mesotrophic nature of the No. 3 seam; b) the groundwater dependence of peatlands as deduced from stratigraphic position; c) evidence of strongly fluctuating flow within channel-belt deposits evidenced by features such as heterolithic form-concordant strata and pebbly sandstone hollow-fills overlying scoured surfaces within the Rodney Sandstone; 4) ephemeral sheetflood recorded in distal alluvial fan deposits; and 5) mire flora (abundant Anabathra and *Lepidodendron* and uncommon *Lepidophloios* and *Densoporites*- producing lycopsids).

It can be stated with some confidence, therefore, that the climate was inequable, and because of paleogeographic reconstruction (Scotese *et al.*, 1979; Rowley *et al*, 1985), tropical. It is possible that the climate was everwet tropical with occasional intense storms as opposed to seasonal wet tropical, however the rheotrophic origin of coals and composition of peatland flora argue against a strictly everwet climate. Furthermore, abundant mottled and red mudrocks in coeval coal-bearing strata of the Joggins section may indicate extended periods of less humid conditions.

As inferred earlier, the orderly development of attributes within the Springhill coalfield basin-fill sequence can be explained solely by improved drainage during basin infilling. A major problem that must be resolved for paleoclimatic reconstruction of the Maritimes Basin is the significance of red mudrocks and associated calcareous nodules: do they represent more arid conditions as widely supposed (Rust *et al.*, 1984; Cecil *et al.*, 1985) or do they simply reflect enhanced drainage associated with basin filling?

Given minimal local orogenic effects (rainshadows) within the Maritimes Basin, a major climatic trend of increased aridity should result in chronostratigraphically constrained red bed sequences between regional depocenters. Unfortunately, the Westphalian B record is largely confined to the Cumberland Basin, therefore there are few coeval strata with which to compare.

It has been illustrated, however, that at least some red beds are associated with progradation of alluvial fan sediments and build-up of the sedimentary surface, leading to a slightly lower groundwater level. These particular red beds therefore may be tectonic signatures of periods of slow or arrested basin subsidence (Blair, 1987; Blair and Biloudeau, 1988). Recent stratigraphic studies (Ryan *et al.*, 1990; Deal, 1990) of the relationship between coarse conglomerates and red beds of the Ragged Reef Formation overlying coal-bearing strata of the Springhill Mines Formation indicate that the red beds are lateral equivalents of the western basin margin conglomerates. The progradation of conglomerates and formation of reddened mudrocks may be attributed to decreased basin subsidence and lower groundwater levels, but may have been enhanced by increasing seasonality (Wilson, 1973; Cecil, 1990).

5.4 CLIMATIC DRIVING MECHANISMS OF CYCLICITY

It is currently popular to embrace astronomically induced (Milankovitch) cycles of climate as the driving mechanism for cyclothems, especially when evidence of transgression-regression is apparent. The Milankovitch climatic cycles arise from three variables of earth orbit and axial orientation: i) orbital eccentricity, with periodicity of 413,000, 95,000, 123,000 and 100,000 years; ii) obliquity (tilt) of the earth axis, with periodicity of 41,000 years; and iii) perihelion precession arising from the circular wobble of the earth's axis, having terms of 23,700, 22,400, 19,160 and 18,980 years (Berger, 1977). The periodicity of cycles in the Carboniferous is thought to have been somewhat shorter, however, due to a progressive increase in earth-moon distance over time (Walker and Zahnle, 1986). Consequently, Collier *et al.* (1990) amended the Late Carboniferous obliquity and perihelion precession-driven cycles to 31,100 years for the obliquity and 20,000, 19,000, 16,500 and 16,700 years (19,500 and 16,600 year average) for perihelion precession.

Eustatic changes in world sea level resulting from the waxing and waning of orbitally forced glaciation are believed to have been responsible for transgressive-regressive cyclothems in the Late Carboniferous (Fischer, 1986; Heckel, 1986), although tectonic overprinting may have been equally important locally (Klein and Willard, 1988). An individual Late Carboniferous transgressive-regressive sequence may comprise 4 or 5 cyclothems (or partial cyclothems), however, especially in areas with abundant clastic input (Heckel, 1986; Ross and Ross, 1988).

Cyclothems of midcontinental USA have been linked (Heckel, 1986) to long cycles of earth orbital eccentricity (ca 400 ka), although subsequent revisions to Carboniferous time scales (e.g. Lippolt and Hess, 1985) lower the calculation of periodicity to 200 ka which does not correlate readily with any known orbital cycle (Klein, 1990). Coal-red bed cycles in the Westphalian D of the Sydney Basin are thought to represent periods of similar duration (220-300 ka; Gibling and Bird, in press). Periodicity of Namurian and Westphalian A-B marine bands (mesothems) in Britain has been calculated at 190 and 200 ka also, although the period may have been as low as 100 ka due to the absence of unequivocally marine strata from much of the Westphalian B (Leeder, 1988). Collier *et al.* (1990) suggested that, on a sloping shelf, some transgressions would reach further inland than others. The further the site from the coast, the greater the likelihood that a transgression event would go unrecorded, thus increasing the apparent period of cyclicity.

In contrast, the time range for cyclothems within the Springhill Mines Formation of 9-15 ka or 32-38 ka is an order of magnitude shorter in duration than that deduced for the Carboniferous of Britain and midcontinental USA. This time frame compares favourably, however, with the time deduced for a cyclothem in Indonesia constrained by C^{14} radiometric dating: the cycle of buried peat to the base of recurring, extant bog forest represents a period of 6 to 32 ka (Cobb *et al.*, in press) which may be an effect of the precession-induced Milankovitch cycle (D. Chesnut, personal communication, 1991).

Eustatic rise and fall of sea level during the Carboniferous (Ross and Ross, 1988), linked to waxing and waning of Gondwanan glaciation (Crowell, 1978) as a result of orbitally forced climatic-change (Heckel, 1986; Chesnut, 1989; Ramsbottom, 1977) is thought to have played a key role in controlling the development of peatlands during the Late Carboniferous. This is especially true, or at least obvious, for coals with obviously marine-influenced cyclothems. From the perspective of a researcher working in the Nova Scotia intermontane coal basins such as the Cumberland Basin for which marine influence has not been documented, however, the influence of such a controlling mechanism is less clear. The presence of bivalve-bearing calcareous shales that commonly overlie coals of the Joggins Formation was interpreted by Duff and Walton (1973) as the signature of increased salinity, perhaps in response to "aborted" marine transgressions. If such were the case, the absence of these fauna within the Springhill Mines Formation suggests that there was even less of a marine influence, or perhaps none at all, during the deposition of these strata. The similarity in time periods of obliquity and precession with the periodicity deduced for the cyclothems of the Springhill coalfield raises the possibility, however, that the Springhill cyclothem was controlled at least in part by climatic change.

5.4.1 Short-term Climatic Driving Mechanisms

Shorter term variations in climate such as ocean upwelling events, sun-spot and lunar cycles may have interposed on longer term variations. The effect of such phenomenon can be dramatic, as shown by the El Nino Southern Oscillation (ENSO) event, which in the early 1980s contributed to major shifts in global precipitation patterns. Drought was experienced in one of the world's most humid areas, Indonesia, with annual rainfall of 3000 mm and may have been largely responsible for devastating forest fires in the peat-forming rainforests of Borneo (East Kalimantan). The so-called "Great Fire of Borneo", thought to have been ignited by lightning strikes which are common in the bog forests (Anderson, 1983), consumed an area in excess of 35,000 km^2 (Johnson, 1984). In contrast, tropical regions of eastern South America experienced century-scale (or greater) floods (Depetris and Kempe, 1990). The ENSO event recurs in 4-5 year period, but is thought to recur with particular severity in 90 year periods (Johnson, 1984). Variations in sediment lamination from Lake Turkana, Kenya over the past 4 ka exhibit a 4 year periodicity, inferred to reflect the ENSO (Halfman and Johnson, 1988). The existence and periodicity of such events in the Paleozoic, however, is unknown to the author.

Although Milankovitch cycles of orbitally induced climatic change have received considerable attention, shorter term astronomically driven agents of climatic change (Fairbridge, 1986) may have had considerable impact on the sedimentologic and coal record as well. Two of these shorter term agents are: i) the 18.6 year lunar nodal cycle, and ii) the 11 and 22-year sunspot cycles. The 18.6 year lunar nodal cycle not only influences tides, but has been identified in tree ring and long term weather records (Currie, 1984). Drought-flood cycles from northern China show a periodicity strikingly similar to the 18.6 year lunar cycle (Currie and Fairbridge, 1985). The 11 year sunspot cycle and 22 year (Hale) cycle in the reversal of the solar magnetic field may also have exerted a short-term control.

It is probable that similar short term agents of climatic change are recorded in major flood-stage deposits within channel belt deposits such as the Rodney Sandstone and within coal seams. Drought stages may correlate at least in part with episodes of major fires (Swetnam and Betancourt, 1990); the record of some of these would have survived as fusain horizons.

5.4.2 Orogenically Induced Climatic Change

One of the most vexing problems in the analysis of the basin-fill record, as stated previously, is whether the upward reddening of the basin fill sequence and concomitant decline in coal reflects decreasing subsidence or a climatic change, or both. The evidence presented in Chapter III suggests that tectonism was a major influence, but this does not rule out a parallel climatic change. If the climate of the basin did experience a change to drier and perhaps stronger seasonal conditions, the time frame would have spanned at least 1 Ma based on available data. Such change cannot be attributed to any known astronomically induced cycle.

Climate change can, however, be induced locally and regionally by orogenic effects. The most obvious such effect would be the creation of a rainshadow for moisture-laden winds. In an equatorial setting, prevailing winds are presumed to have been easterly (from the east). Until the exact paleogeographic orientation of basins and intervening mountain ranges within the parent Maritimes Basin has been deduced, however, such effects remain hypothetical.

The minimum elevation at which orogenically induced climatic change is thought to be possible ranges from 500 m to as much as 3000 m depending on relative humidity and temperature (Hay *et al.*, 1982; Perlmutter and Matthews, 1989). A tectonically induced cycle of climate change, encompassing orogenic uplift and denudation, would span a range of 610,000 to 10.5 m.y. or longer without isostatic compensation (Perlmutter and Matthews, 1989). Their calculations were based on a maximum erosion rate of 90 cm/yr. for a tropical, very humid environment (Leopold *et al.*, 1964) and rates of uplift of 5 cm/kyr for rift margins (Seidler and Jacoby, 1981) to as great as 800 cm/kyr in major orogenic belts (Schumm, 1963). It follows, therefore, that orogenically induced climatic change was possible at least from a temporal standpoint during deposition of the basin-fill sequence. Red bed sequences in the Late Carboniferous of the Pennine Basin, U.K., have been attributed to such local tectonic effects (Besly, 1988).

5.5 CLIMATE CHANGE AND THE DEVELOPMENT OF CONTINENTAL TROPICAL MIRES AND CYCLOTHEMS

The time span deduced for a continental Springhill cyclothem, in the order of 10-40 ka, approximates that of the precession and possibly obliquity-induced Milankovitch cycles. It follows, therefore, that climate change probably occurred during cyclothem formation. The origin of marine-influenced or near coastal cyclothems can be readily explained by climatically induced changes in sea level (Wanless and Shepard, 1936; Cobb *et al.*, 1989; Gibling and Bird, in press). The role of climate change in the development of continental cyclothems is much less obvious.

Climate change resulting from Milankovitch cycles would have been most pronounced in mid-latitude regions, with the least change occurring in regions of climatic extremes, these being the tropics and poles (Perlmutter and Matthews, 1989). Paleogeographic reconstruction (Scotese *et al.*, 1979; Rowley *et al.*, 1985) places the Cumberland Basin in a near equatorial position during the Westphalian B. Modelling of climatic change during Milankovitch cycles suggests that in areas of less than 15° latitude, a tropical climate would have prevailed even during the climatic minimum (Perlmutter and Matthews, 1989). Decrease in area of the Hadley cell of atmospheric circulation and narrowing of the Intertropical Convergence Zone (ITCZ) during the climatic minimum, however, would result in less humid conditions and increased seasonality in all but the lowest of tropical latitudes (Permutter and Matthews, 1989). Accompanying this change would be a much decreased monsoonal circulation (Fairbridge, 1986), an important source of moisture in many tropical regions.

5.5.1 Impact of Climate Change on Tropical Mires

The precise effect of climate change on mire development is by no means well understood. As stated by Frenzl (1983, p. 44), "the most elusive question concerns climatic control of peat growth". Nonetheless, climate change has the potential to alter the water balance of a mire, and in so doing to effect profound change in its development. The water balance equation, discussed previously in Chapter III, was given by Ingram (1983) as:

$$P + N - D - E - \Delta W - n = 0$$

[precipitation (P) + subsurface discharge (N) in balance with discharge (D) + evapotranspiration (E) and water storage (ΔW), n representing error]

Simply stated, recharge to the mire equals discharge (Moore, 1987).

Precipitation is perhaps the most obvious variable affected by climate change. Rather than change in the annual volume of precipitation, however, the seasonal distribution of rainfall is of far greater importance to the development of tropical mires (Thompson and Hamilton, 1983). This is especially true for peat-forming ecosystems in tropical continental settings without benefit of a humid maritime climate. Of similar importance is change in temperature. Elevated temperature will not only affect evapotranspiration, hence water balance in the mire, but the rate of decomposition, hence peat accumulation.

The effects of seasonality and the relation between precipitation and temperature are well illustrated by the distribution and histories of continental African mires. Although vast permanent and seasonally inundated herbaceous and forest wetlands occur over many areas (in the central basin of the Zaire River, as much as 80,000 km^2), only a small percentage of these ecosystems are peat-forming (Thompson and Hamilton, 1983). Those which are peat-forming can do so in the face of unfavourable seasonality either because 1) they are at sufficient altitude to reduce temperature and thereby reduce evapotranspiration and decomposition, or 2) they benefit from groundwater systems that supply recharge during the dry season. In Africa, solely rainfed ombrogenous peats nowhere form at altitudes of less than 2000 m (Thompson and Hamilton, *ibid.*)

Thompson and Hamilton described the histories of several African mires, constrained by radiocarbon dating, against the backdrop of Late Pleistocene-Holocene climate change. The four mires discussed by Thompson and Hamilton are: i) Kamiranzovu Swamp, Rwanda ("relatively low altitude"); ii) Karimu Swamp, Kenya (ca. 3000 m elev.); iii) Kaisungor Swamp, Charangani, Kenya (2900 m elev.); and iv) swamps of Mt. Badda, Ethiopia (4040m elevation). Three quite different responses are recorded, depending in large part on the altitude of each deposit. The history of central African climate during the past 35 ka can be summarized as follows:

35,000 to 22,000-20,000 B.P.: cool, moist
22,000-20,000 to 12,500 B.P.: cool, arid
12,500 to 10,000 B.P.: transitional
10,000 to 4,000 B.P.: moist
4,000 B.P. to present: drier

Kamiranzovu Swamp is a very slightly domed mire in a moist climatic setting. Peat has accumulated here over a period of 25-27 ka, from ca. 37,630 to 12,600-11,000 B.P., however, the ecosystem has achieved equilibrium and has stagnated since then, with no appreciable peat accumulation for at least 11 ka. Importantly, peat accumulation ceased during a period of increased precipitation: the period from 12,500 to 10,000 yrs. B.P. saw the level of nearly Lake Kivu rise from 300 m below its present elevation to achieve outflow. Thompson and Hamilton (p. 361, 1983) postulated that "a tendency for peat growth to be encouraged by increased waterlogging was more than offset by higher temperatures." A hiatus in peat accumulation during this moist period from 10-4 ka B.P. is also recorded in the high altitude mires of Charangi, Kenya and Mt. Badda, Ethiopia. Thompson and Hamilton suggest that peat accumulation in these mires was interrupted by increased flow. Thus, although peat accumulation ceased during the same period in both settings, the reasons differed.

In contrast, up to 2.5 m of peat has accumulated in Karimu Swamp, Kenya during the last 9150 years. Mire growth occurred here during moister times, with inorganic sedimentation during drier times when vegetative cover was reduced. Peat accumulation was terminated then, by elevated temperature in the lower altitudes, and in higher altitudes by drier conditions resulting in increased runoff or by moist conditions resulting in excessive flow. The response of ancient mires will have varied, therefore, with geographic setting. Furthermore, one cannot assume that in all cases continental peat accumulation was favoured during warm periods of climate maxima/interglacials.

5.5.2 Implications for Springhill Paleomires and Cyclothems

Peat accumulation in tropical Africa has been favoured during climatic periods that are a) less seasonal; b) moist; and c) cool. Cecil (1990) advanced a climate-induced model for cyclic sedimentation that considers precipitation/humidity but assumes a continually warm climate. In this model, coal (peat) represents a moist, minimally seasonal climate, siliclastics represent strongly seasonal climate and pedogenic carbonates and evaporites represent the most arid, minimally seasonal conditions. Applied to the Springhill cyclothem, coal would represent a climatic maximum with equally moist conditions, and multistorey sandstone would represent strongly seasonal conditions. Arid conditions may not have accompanied climate change in an equatorial setting, however, according to the climate change model of Perlmutter and Matthews (1989).

The history of some tropical mires of the African continent, however, illustrates the inadequacies of such simple models. If the Springhill paleomires did not benefit from a moist maritime climate as in modern Southeast Asia, for example, the climatic optimum may have been less favourable for peat accumulation than was the climatic minimum. Future research into the Westphalian paleoclimate and paleogeography of the Maritimes Basin and Euramerica may help to clarify the application of a climate change model in the development of cyclothems of the Springhill coalfield. It is likely, however, that paleoclimate was influential, even if the precise mechanisms are not clear. Clearly, any factors that would result in decreased rate or especially stagnation of peat accumulation in the rapidly subsiding Cumberland Basin would have rendered the paleomire susceptible to burial.

5.6 TECTONIC CONTROLS

The major influence of tectonism on peat accumulation and preservation was the provision of a suitable basin and paleogeographic setting. This included an environment wherein peatlands could co-exist with groundwater-supplying piedmont fans, generally in retreat, and axial rivers, presumably in an extensional or transtensional regime. An omnipresent, limiting factor was the rapid subsidence rate experienced by the basin, with approximately 4 km of lithified strata deposited largely during the Westphalian A & B (4 Ma; Lippolt and Hess, 1985; Hess and Lippolt, 1985). In the Athol Syncline, a time period of 1 million years is represented by 650 to 1000 m of strata; in comparison, a similar time span in the Sydney Basin (Sydney Mines Formation: Westphalian D) is represented by a mere 150 m (Gibling and Bird, in press). Although long-term (10^6 yr.) climatic change may have accompanied basin infilling, the basin-fill sequence and coal window can be attributed to a progressive decline in net subsidence rate, largely influenced by tectonism. Similarly, tectonic readjustment of the basin was likely responsible for red bed-conglomerate events within the basin fill.

It would seem illogical, however, to attribute the development of the coal-mudrock-multistorey sandstone cyclothem to tectonism alone. The regular cyclicity and thickness of these deposits imply a cyclic temporal control not readily explained by random fault readjustments (earthquakes). Local topographic control on channel belt and mire position (e.g. Nos. 1 and 3 seams), however, may well have been fault controlled. It may be possible that tectonism acted as a triggering mechanism, inducing a cycle of channel-belt and mire processes that, because of autogenic evolution, would require a certain time span to unfold.

5.6.1 Diapirism

One of the potentially most important controls of the basin-fill sequence scale was salt diapirism. Salt diapirism has been suggested as a fundamental control on coal formation in the Springhill coalfield (Hacquebard *et al.*, 1967). The deterioration of coals across the present day anticlinal axis, however, earlier interpreted as evidence of positive topographic relief induced by diapirism, has since been shown to be the result of progressive onlap of seams toward the basin margin and interaction with basin margin deposits (Calder, 1981b; 1981c; 1986; and this study). Diapirism of Windsor Group evaporites is thought, however, to have occurred prior to the late Westphalian (Bell, 1944; Ryan, 1986) and salt kinetics may have played an important role in providing accommodation space for the accumulating peat-bearing sediments. The recovery of reworked Visean (Windsor Group age) spores from the riverine zone of the No. 3 seam (Dolby, 1988a) is evidence that the Windsor Group was being eroded in the hinterland during accumulation of the ancient Springhill peats.

5.7 TOWARD AN INTEGRATED CLIMATIC - SUBSIDENCE MODEL OF CYCLOTHEM DYNAMICS

The following 'model' briefly describes the history of a continental Springhill cyclothem, identifying likely controls or influence at various stages of its development. For the purpose of discussion, the development of a Springhill cyclothem can be considered as comprising the following stages:

1) An emergent groundwater table fed by nutrient-rich seepage from distal alluvial fans, and quite possibly from related basin margin springs, provided a site conducive to luxuriant plant growth. Waterlogging of the soil (paludification) would have been enhanced by pedogenic iron carbonate formation in the substrate (Frenzl, 1983).

2) Peat accumulates for thousands of years beneath a forest cover dominated by arboreous lycopsids and nourished by distal alluvial fan discharge. During this fertile stage of mire development, peat accumulation rates equal or exceed basin subsidence. The climate is inferred to have been humid and relatively if not wholly equable but may have been either warm or cool.

3) Peat accumulation slows due to either decreasing fertility (cf. Anderson, 1983) or water supply. The supply of water to the ecosystem may have been affected by climate stress in the form of increased temperature or seasonality (cf. Thompson and Hamilton, 1983), or by other causes such as eustatic draw-down of the regional water table (see Summary). Basin subsidence rates continue unabated.

4) The mire reaches equilibrium.

5) Subsidence outstrips peat accumulation and the mire is buried. Compounding this imbalance may have been increased siliciclastic deposition arising from climate change (seasonality; Cecil, 1990). The mire would also have been particularly susceptible to tectonism at this stage. Autocompaction of peat would have accentuated subsidence of the mire.

6) Mud accumulates in perennially submerged floodplain lakes over the compacting peat; local thin limestones accumulated at sites or during period of sediment-starvation. The topographic low provides a site for river avulsion, triggered either by autocyclic processes or tectonism, and possibly promoted by increased sedimentation in the channel-belt enhanced by a seasonal climate (cf. Cecil, 1990).

7) After accreting to a height susceptible to either autocyclically or allocyclically induced avulsion, the river returns to a basin axial position, leaving an abandoned, nutrient-rich river tract and floodplain.

8) Peat accumulation resumes after repopulation by wetland flora dominated by arboreous lycopsids. Peat accumulation may have benefitted from decreasing seasonality.

Any model of cyclothem development at Springhill, however, must address the occurrence of "in-seam" cyclothems, such as the riverine zone split in the No. 3 seam (Figure 4.3). A cyclothem-like sequence of coal, mudrock and multistorey sandstone occurs within the riverine zone of the No. 3 seam, but laterally in the inner mire zone the seam survived intact. It is difficult, therefore, to invoke a wholly climate-induced mechanism for cyclothem development. Even if the period of siliciclastic deposition in the riverine zone were to correspond with stagnation of the mire, (a hiatus), in the inner mire zone, there must have been some factor that accounted for the different depositional histories of the two zones. The most plausible are local fault control (a fault parallels the trend of the channel-belt deposits entombed within the No. 3 seam, although the timing of movement is unknown) or differential compaction and nesting of sandstone bodies (the inner mire zone may sit atop an underlying multistorey sandstone body).

Evidence of allogenic environmental change, coincidental with the development of the riverine zone 'cyclothem' is found in the apparently simultaneous deepest incursion of distal fan sheetflood deposits in the piedmont zone, which together almost succeeded in overwhelming the mire. A dramatic decline in *Anabathra*, represented by *Lycospora orbicula*, occurred within the inner mire at this time, (Figure 4.12) but it is not clear whether this floral change resulted from tectonic, climate or even autogenic change. It has been noted, however, that this change occurs above a fusain horizon, product of wildfire.

5.8 SUMMARY

When one considers the present, it seems inconceivable that one would not recognize the obvious climatic control on zonation of plants. Basins, the receptacles of peat accumulated from certain of these floras, on the other hand, are invariably a tectonic phenomenon (with the possible exception of glaciated valleys of short geologic life). The interaction of climatic and basin-forming mechanisms may be especially important in the case of rheotrophic peatlands which, under marginal climatic conditions, are necessarily dependent upon basin groundwater to support vegetation and retard decomposition.

Yet in geological studies of ancient peat formation, one of these controls (climate or tectonism) is routinely embraced while the other is similarly discounted or ignored. In intermontane basins of limited areal extent in which tectonic effects are readily apparent, tectonism is usually embraced (e.g. Courel *et al.*, 1986; Naylor *et al.*, 1989). In basins of widespread extent in which basin margin deposits are less represented, climatic or eustatic effects tend to be embraced (e.g. Wanless and Shepard, 1936; Heckel, 1986; Cecil, 1990) and perhaps this is correct. In the Springhill coalfield, both tectonism and climate were influential, but the effects of tectonism are more obvious.

It is clear that *no one control can be cited as having been solely responsible for peat accumulation and coal formation in the Springhill coalfield, (nor probably in any coalfield)*, and there is evidence that controls interacted, such as the long term tectonic and climatic framework. The primary controls on coal formation had to allow for *establishment* of peatlands, as well as their *longevity*. These included a suitable climate and groundwater-supplying deposystems from which peatlands could be nourished and with which peatlands could co-exist. For the peats to have been preserved, accommodation space for the accumulating sediments had to be provided.

A hierarchy of controls operated (Figure 5.3): some interacted to form a compound control, whereas others interposed on those of longer duration. Two fundamentally important levels of control operated: one resulting in the *coal window* within the basin-fill sequence, possibly in the order of 10^6 yrs.; and another at the *cyclothem* level, possibly in the order of 10^4 yrs. The coal window may have been constrained by net subsidence, potentially in concert with orogenically induced climatic change. Denudation of the Cobequid highlands (Ryan *et al.*, 1987) would have resulted in diminished groundwater supply through a reduction in catchment area and fan size (Denny, 1965; Hunt, 1975) and possibly loss of orographic rainshadow. The cyclothem most probably reflects autogenic mire and fluvial processes operating in part under constraint of orbitally-forced climatic(/ geoidal?) cycles of obliquity or precession and basin subsidence. Sole tectonic control was unlikely due to the well ordered periodicity, but tectonism could have played a role in triggering the cyclothem, after which natural autogenic processes (e.g. channel-belt migration and mire genesis) took their course.

Still further controls operated in the year to centuries and even millennia range during the course of a mire's life. Floods, probably responsible for high ash bands within coal lithotypes and for some impure coal lithotypes, probably occurred in the year to decades (cf. Gupta, 1983) to century (cf. Depetris and Kempe, 1990) range. Especially severe floods are also recorded within fluvial channel belt deposits. Fires, represented by mm-scale fusain lenses and occasional cm-thick layers were, as today, ongoing agents of modification in the peat-forming ecosystem. Such events may have been especially severe at the century scale, although more detailed scrutiny of fusain distribution at the microscopic level (e.g. polished blocks) may show evidence of a much higher frequency. These short term events may be tied to phenomena such as the lunar nodal cycle and to non-orbitally related climatic fluctuation related to phenomena such as oceanic upwelling and sunspot cycles. Such short term driving mechanisms, which may interact to reinforce or cancel their effects, are as yet only partially understood. Autogenic mire evolution, especially floral succession, may have operated at the millennia scale over the life of the mire, but may have been less a contributing factor to ultimate seam thickness and cyclothem dynamics than were climate change and subsidence.

Processes of mire development in the Springhill coalfield operated within the constraints of basin processes, although the two sometimes interacted through processes such as lateral paludification and distal fan sheetflow. Both levels of control were affected, however, by short and

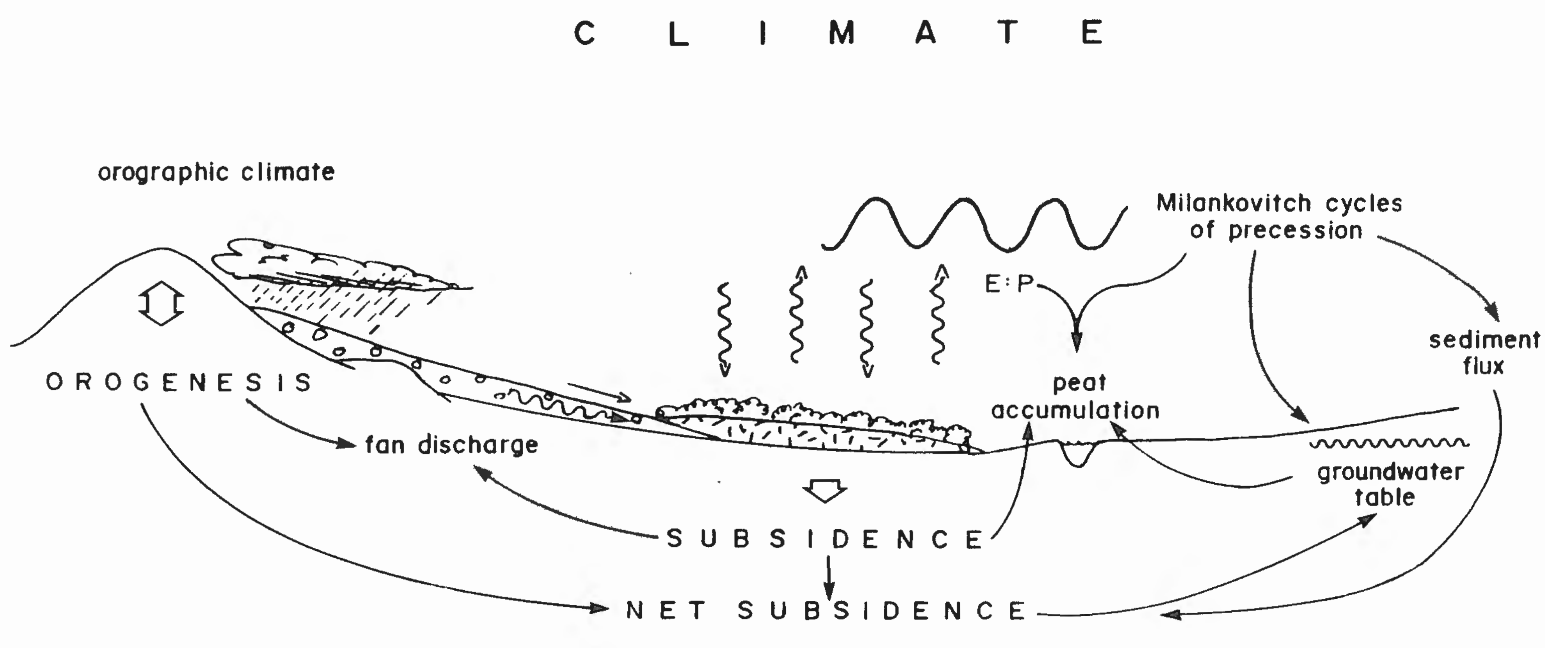


Figure 5.3 Schematic interaction of major processes that affected Westphalian peat accumulation at Springhill.

long term climatic and by orogenic processes which in turn had potential to induce long-term climatic change.

Tectonically related recharge to peat-forming systems is less important under optimal climatic conditions of low evapotranspiration: precipitation and seaonsality than under less favourable climatic conditions (Figure 5.4). The observation by Bower (1961) with respect to blanket peats of the Pennines may be appropriate: "One could say that the climate determines the general limits within which landform determines the details".

The similarity of basin-fill sequences and seam geometry of the Springhill coalfield with certain other intermontane basins such as the Mesozoic Fuxin Basin, China, Pennsylvanian Anthracite Basin of Pennsylvania and late Westphalian-Stephanian basins of Spain suggest that some of these controls may be applicable to other basins, although no two basins will be precisely identical. Intermontane peatlands associated with the Variscan Cantabrian mountain belt in northern Spain formed in the late Westphalian and early Stephanian in the face of widespread redbed formation in Europe (Bless *et al.*, 1981). Besley (1988) suggested that the Cantabrian peats were beneficiaries of an orogenic rainshadow because of their intermontane location. Their proximity to alluvial fan systems (Heward, 1978) suggests, however, that fan discharge may have been important to their development.

The location and stratigraphic position of coal within the Cumberland Basin, of primary importance to exploration, was largely affected by basin subsidence history and by paleogeographic effects on groundwater distribution. So strong is the evidence in support of mires nourished by groundwater from alluvial fans, that it is unlikely that thick coals will be found in the Cumberland Basin far removed from these ancient landforms. This is especially true within the younger parts of the basin-fill sequence, when groundwater is inferred to have progressively diminished. The physical mineability of the coal resource will be affected by its paleogeographic setting, by the effects of differential compaction, and perhaps by local fault control, which impact on geometric zonation of the seams.

Much of the character of a seam that affects its desirability as an energy or mineral resource including sulphur, ash and other mineral content is a direct result of its trophic and topographic status (rheotrophic, planar vs ombrotrophic, raised), which is dependent in large part upon climate (distribution and total of annual precipitation and the precipitation/evaporation ratio). As a direct result, rheotrophic (planar to weakly domed) mires will be susceptible to a much higher input of

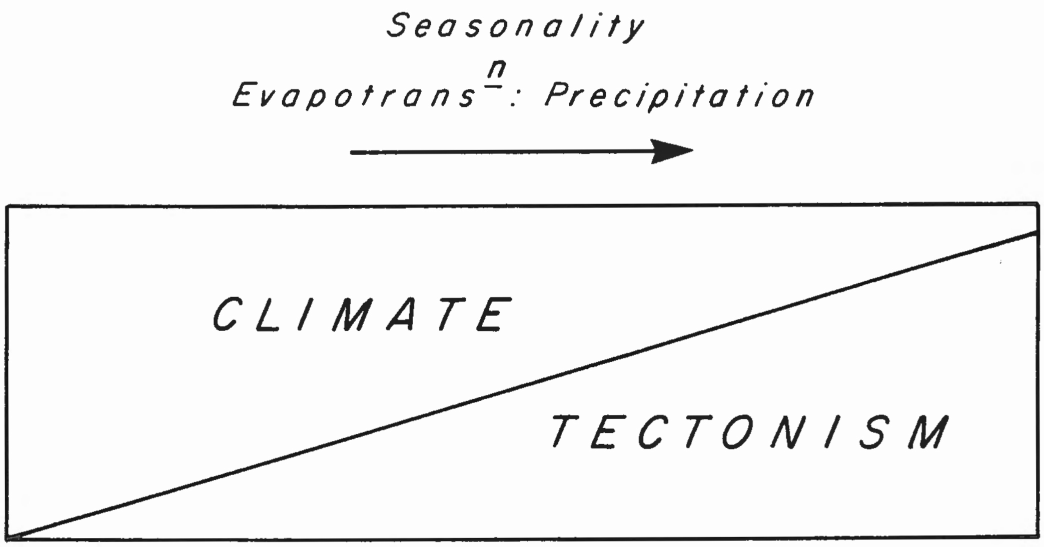


Figure 5.4 A schematic depiction of the relative effects of climate and tectonism on basin hydrology. As climate becomes less optimal, the importance to wetland development of tectonically derived recharge increases.

mineral matter and ions from neighbouring deposystems. The chemistry of rheotrophic ecosystems therefore will have been profoundly affected by the source of groundwater, which in the Springhill coalfield differed between piedmont and riverine zones. The rapid subsidence history of the basin may have contributed to the rheotrophic character of the mires, but without an adequate balance of precipitation to evaporation, a truly domed status would not have been possible.

The refinement of controls on coal formation requires further systematic, comprehensive studies which test objective criteria based on modern peatland ecology modified to take into account the different biological character of Late Carboniferous flora and set within the framework of regional geologic processes. Our understanding of these controls will continue to be constrained in large part, however, by our knowledge of the climatic and temporal frameworks of ancient peat accumulation.

5.8.1 Recommended Areas of Study: Further Tugs at the Gordian Knot

This research has perhaps outlined as many areas that are fundamentally important yet enigmatic as it has those that are clearly understood. Some areas of recommended future study that will continue to unravel the Gordian Knot of ancient mire ecosystem dynamics are:

- The hydrogeology of basins that contain major peat-forming ecosystems, and especially the study of modern peats associated with alluvial fans.

- Study of pathways of degradation in peat and resolution of the effects of plant type vs. mire chemistry which may require changes in the prevailing nomenclature of vitrinite macerals. Such research will permit a much clearer understanding of mire genesis.

- The origin of continental cyclothems: specific areas to be investigated include the effects on inland basins of eustatic change and changes in the geoid. Mörner (1984, p. 412) stated "geoid lowering under the continents leads to a corresponding groundwater lowering... that may have drastic effects on the flora and fauna". His correlation of geoidal changes with Milankovitch cycles exhibiting a similar periodicity to the cyclothems of the Springhill coalfield merits further investigation.

- The paleoecology of ancient mires elsewhere in the Maritimes Basin has been little studied. This study has shown that the flora are accurate reflections of the ancient environment and should therefore be incorporated in future paleoenvironmental studies of coal seams.

- Comprehensive assessment of Carboniferous paleoclimate throughout the Maritimes Basin will be fundamentally important in constraining future studies of the origin and distribution of coal deposits in the area, but will require careful consideration of the validity of paleoclimate indicators. This will also permit a clearer understanding of the overall development of Euramerican paleoclimate during the Carboniferous.

- For the Cumberland Basin, tectonic analysis of the origin of the basin is fundamental to constraining future models of coal exploration. Analysis of the Athol-Sand Cove fault zone and its regional implications is especially recommended.

APPENDICES

A. Relative abundance of lithofacies for various measured sections of lithofacies assemblages I-VI.

B. Detailed macroscopic logs of selected seam profiles.

C. Maceral and chemical analyses, No. 3 seam.

LITHOFACIES	TOTAL THICKNESS (m)	n	% of EXPOSED SECTION	LITHOFACIES GROUP PERCENTAGE
Gt	9.49	9	34.8	
Gm	9.66	8	35.5	G = 71.4
Gg	0.29	1	1.1	

St	3.59	8	13.2	
Sg	1.80	1	6.6	
Sr	0.85	2	3.1	
Sh	0.51	2	1.9	S = 26.4
Sm	0.26	1	0.9	
Ss	0.20	2	0.7	

Fm	0.60	1	2.2	F = 2.2

C	-	-	-	Nil

L	-	-	-	Nil

Table A-1 Summary of lithofacies abundance in conglomerate assemblage (I), Polly Brook.

LITHOFACIES	TOTAL THICKNESS (m)	n	% of SECTION	LITHOFACIES GROUP PERCENTAGE
Gm	1.52	1	3.9	G = 3.9
- - - - -				
Sg	6.20	13	15.9	
Si	4.12	2	10.6	
Sm	1.32	2	3.4	S = 32.2
Sd(c)	0.41	1	1.1	
Ss	0.35	1	0.9	
Sr	0.12	1	0.3	
- - - - -				
Fi	15.91	5	40.9	
Fm	4.34	5	11.2	
Fl	2.71	7	7.0	F = 62.3
Fc	1.11	3	2.9	
Fg	0.12	1	0.3	
- - - - -				
Ci	0.21	3	0.5	C = 0.5
- - - - -				
Lc	0.31	3	0.8	L = 1.2
Lm	0.15	1	0.	

Table A-2 Summary of lithofacies abundance for poorly sorted lithofacies assemblage (II), drillhole SH38, 120.11 - 122.0 m and 124.0 - 162.20 m.

LITHOFACIES	TOTAL THICKNESS (m)	n	% of EXPOSED SECTION	LITHOFACIES GROUP PERCENTAGE
Gg	0.46		6.9	G = 11.4
Gm	0.30		4.5	
- - - - - -				
Si	3.30		49.2	
Sl	0.90		13.4	
St	0.60		8.9	S = 82.9
Sm	0.35		5.2	
Sh	0.24		3.6	
Sg	0.18		2.7	
- - - - - -				
Fm	0.38		5.7	F = 5.7
- - - - - -				
C	-	-	-	Nil
- - - - - -				
L	-	-	-	Nil

Table A-3 Summary of lithofacies abundance in exposure of poorly sorted assemblage (II) on Wyndham Hill road.

LITHOFACIES	TOTAL THICKNESS (m)	n	% OF EXPOSED SECTION	LITHOFACIES GROUP PERCENTAGE
G				G = Nil
- - - - -	-	-	-	
Sd	11.55	22	7.18	
Shl	10.29	17	6.40	
Si	8.20	6	5.10	
Sm	5.77	11	3.59	
Sg	4.87	7	3.03	S = 29.1
Sh	3.71	6	2.31	
Se	1.18	1	0.73	
Sr	0.89	2	0.55	
Ss	0.30	1	0.19	
- - - - -				
Fi	27.38	12	17.02	
Fm	25.59	35	15.91	
Fcl	25.58	25	15.90	
Fl	12.74	23	7.92	F = 67.8
Fc	10.44	20	6.49	
Flrh	4.61	2	2.87	
Fg	2.67	3	1.66	
- - - - -				
Lc	1.18	3	0.73	
Lm	0.68	6	0.42	L = 1.9
Lcc	0.66	1	0.41	
Lb	0.61	1	0.38	
- - - - -				
Csi	0.93	4	0.58	
Ci	0.90	8	0.56	C = 1.2
Cd	0.11	1	0.07	

Table A-4 Summary of lithofacies abundance in thinly interbedded assemblage (III), drillhole SH74 (100.08 - 224.31 and 233.44 - 282.78 m)

LITHOFACIES	TOTAL THICKNESS IN SECTION (m)	n	% OF MEASURED SECTION	LITHOFACIES GROUP PERCENTAGE
G	-	-	Nil	G = 0

Shrh	8.14	5	38.71	
Sr	1.65	2	7.85	
Sm	0.83	2	3.95	
Si	0.33	1	1.57	S = 57.3
Sdb	0.32	1	1.52	
Shl	0.32	1	1.52	
Sh	0.30	1	1.43	
Sg	0.15	1	0.71	

Fm	3.13	5	14.88	
Fl	0.92	2	4.37	F = 22.1
Fb	0.92	1	1.38	
Fc	0.20	2	0.95	
Fi	0.10	1	0.48	

Lb	3.07	5	14.60	L = 14.6

Csi	0.71	2	3.38	
Ci	0.54	2	2.57	C = 6.1
Cv	0.03	1	0.14	

Table A-5 Relative abundance of lithofacies within bivalve-bearing assemblage (IIIBv).

LITHOFACIES	TOTAL THICKNESS (m) SH 5; 312.6-479.0m	SH 99; 22.18-156.65m	SH 74; 10.06-100.08m	THICKNESS (t) AND PERCENTAGE (%) OF COMPOSITE SECTION t	%	LITHOFACIES GROUP PERCENTAGE
G	-	-	-	-	-	Nil
Si	9.09	9.98	1.90	20.97	6.00	
Sd	12.79	10.49	2.48	25.76	7.37	
Sm	16.37	1.71	2.45	20.53	5.87	
Sr	9.10	8.17	3.17	20.44	5.85	S = 36.1
Shl	5.96	7.73	0.46	14.15	4.05	
Se	12.46	0.68	0.39	13.53	3.87	
Sh	4.42	3.82	1.40	9.64	2.76	
St	-	1.28	-	1.28	3.67	
Fm	36.66	17.99	28.07	82.72	23.65	
Fc	22.18	10.36	19.24	51.78	14.81	
Fl	5.95	16.76	3.95	26.66	7.62	F = 60.9
Fcl	6.68	19.07	-	25.75	7.37	
Fi	7.99	7.03	0.67	15.69	4.49	
Flrh	1.01	-	-	1.01	2.89	
Fcs	-	-	0.37	0.37	0.11	
Ci	2.55	5.54	2.56	10.65	3.05	
Cc	-	2.62	0.51	3.13	0.90	
Cd	0.06	1.25	1.60	2.91	0.83	C = 4.9
Cv	-	0.34	-	0.34	0.10	
Cf	-	0.11	0.01	0.12	0.03(tr)	
Lcc	0.37	-	0.71	1.08	0.31	
Lm	0.50	0.30	0.10	0.90	0.26	L = 0.7
Ls	0.21	-	-	0.21	0.06	
Lc	-	0.08	-	0.08	0.02	

Table A-6 Relative abundance of lithofacies in composite reference section (SH5, SH99, SH74) of mudrock/multistorey sandstone assemblage.

LITHOFACIES	TOTAL THICKNESS (m)	n	% of SECTION	LITHOFACIES GROUP PERCENTAGE
G	-	-	Nil	G = Nil

Sh	22.35	13	23.92	
Sm	11.68	8	12.50	
Se	7.47	11	8.00	
Sg	4.05	3	4.34	S = 57.6
Sd	3.81	4	4.08	
Sr	1.80	3	1.93	
Si	1.45	2	1.55	
Shl	1.22	1	1.31	

Fm	33.57	25	35.93	
Fl	5.23	2	5.60	F = 42.0
Fi	0.46	1	0.50	

Lcw	0.27	1	0.29	L = 0.4
Lcc	0.06	1	0.06 (tr)	

Table A-7 Relative abundance of lithofacies within red mudrock/lithic-sandstone assemblage (V); drillhole SH5, 3.35-143.90 m.

LITHOFACIES	TOTAL THICKNESS (m)	n	PERCENTAGE OF TOTAL SECTION	LITHOFACIES GROUP PERCENTAGE
G	-	-	Nil	G = Nil

Sh	29.64	18	19.73	
Sm	10.47	14	6.97	
Shl	9.59	9	6.38	
Sdb	6.79	6	4.52	
Si	6.55	3	4.36	S = 49.9
Sr	5.85	8	3.89	
Se	3.59	14	2.39	
St	3.56	1	2.37	
Sdc	1.28	2	0.85	
Ss	1.19	3	0.79	

Fm	57.09	50	38.00	
Fl	7.65	12	5.09	
Fi	5.10	5	3.39	F = 49.8
Fc	2.97	5	1.98	
Fcl	2.01	1	1.34	

C	-	-	-	C = Nil

Lp	0.34	2	0.23	L = 0.3
Lm	0.12	1	0.08	

Table A-8 Relative abundance of lithofacies in variegated mudrock/multistorey sandstone assemblage (VI); drillhole SH5, 143.90-312.60 m.

MACROSCOPIC PROFILE
SEAM: No. 3
LOCATION: Drillhole SH72 (RIVERINE ZONE)

SAMPLE PETROG./ PALYN.	THICKNESS(m)		LITHOTYPE/LITHOFACIES
1UT	40.18-40.22	-	highly impure coaly shale
2UT	40.22-40.28	-	laminated carbonaceous mudrock (Fc)
	40.28-40.58	-	lost core (coal)
3UT	40.58-40.61	-	shaly coal, microbanded
4UT	40.61-41.00	-	dark grey, slickensided, slightly carbonaceous mudstone
5UT	41.00-41.03	-	black, coaly shale to coaly mudstone with thin vitrain bands
5UT	41.03-41.10	-	intermediate to semi-bright microbanded duroclarain; trace of fusain on bedding
6UT	41.10-41.15	-	bright to semi-bright microbanded clarain (in places grading to duroclarain)
7UT	41.15-41.21	-	micro-banded clarodurain with few mm thick carbonaceous mudstone laminae
7UT	41.21-41.215	-	fusain
8UT	41.215-41.27	-	semibright clarain with fusain lenses
9UT	41.27-41.32	-	intermediate clarodurain to duroclarain
9UT	41.32-41.34	-	thickly banded clarodurain and carbonaceous mudstone
10UT	41.34-41.43	-	semi-bright, micro-banded to normal banded clarain
11UT	41.43-41.47	-	dull duroclarain with several mm thick carbonaceous mudstone lenses

SAMPLE PETROG./ PALYN.	THICKNESS(m)		LITHOTYPE/LITHOFACIES
	41.47+	-	(seat earth rooted to at least 1m.)
			Lower Leaf:
1LT	64.48-64.52m	-	thinly banded, black carbonaceous to coaly shale
1LT	64.52-64.62	-	microbanded dull shaly durain
2LT	64.62-64.75	-	dark grey, carbonaceous mudstone (Fc)
2LT	64.75-65.15	-	dark grey carbonaceous mudstone, coarsening upward to silty mudstone
3LT	65.15-65.18	-	dull black coaly shale with vitrain microbands
4LT	65.18-65.34	-	dark grey mudstone, carbonaceous at top (0.10m recovered)
5LT	65.34-65.42	-	microbanded coaly shale
6LT	65.42-65.47	-	dull, microbanded clarodurain
6LT	65.47-65.52	-	dull, microbanded shaly durain
7LT	65.52-65.59	-	dark grey carbonaceous mudstone (Fc)
8LT	65.59-65.61	-	dull, micro-banded, shaly durain
9LT	65-61-65-63	-	dark grey to black, carbonaceous mudstone (Fc)
9LT	65.63-66.23	-	dark grey mudstone with common Calamites sp. and rare diffuse siderite lenses
10LT	66.23-66.28	-	dull micro-banded and occasionally normal-banded shaly coal (durain)

SAMPLE PETROG./ PALYN.	THICKNESS(m)		LITHOTYPE/LITHOFACIES
11LT	66.28-66.34	-	black, thinly laminated coaly shale
11LT	66.34-66.37	-	dull shaly coal
11LT	66.37-66.43	-	dull, microbanded shaly durain
12LT	66.43-66.86	-	dark grey, wavy laminated carbonaceous mudstone; 2 cm dull black carbonaceous lamina at 66.60m
13LT	66.86-66.88	-	dull black carbonaceous mudstone
	66.88-	-	lost core (coal)

MACROSCOPIC PROFILE
SEAM: No. 3
LOCATION: Drillhole SH81 (INNER MIRE ZONES)

SAMPLE PETROG./PALYN	THICKNESS(m)		LITHOTYPE/LITHOFACIES
	0.23	-	lost core
	0.03	-	micro- and normal-banded clarain
SH81-1T	0.03	-	normal banded clarain
	0.01	-	normal banded duroclarain and clarain
	0.08	-	clarain with pyrite on cleats
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	0.01	-	pyrite lens in fusain
	0.01	-	semi-dull clarodurain
	0.01	-	semi-bright clarain
SH81-2T	0.08	-	semi-bright duroclarain with regularly spaced <1mm clarovitrain bands every 3-4mm.
	0.06	-	clarain with numerous 2mm fusain bands and lenses throughout
	0.07	-	duroclarain with occasional <1mm ferruginous lenses
	0.01	-	semi-dull clarodurain
	0.04	-	bright clarovitrain
	0.01	-	semi-dull clarodurain
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	0.29	-	lost core
SH81-3T	0.09	-	bright vitroclarain

SAMPLE PETROG./PALYN	THICKNESS(m)		LITHOTYPE/LITHOFACIES
	0.06	-	semi-bright clarain
	0.02	-	dull banded durain with <1mm vitrain bands
	0.02	-	intermediate duroclarain
SH81-3T	0.04	-	semi-dull clarodurain with micro-banded fusain in upper 1cm
	0.04	-	intermediate to semi-bright clarain
	0.03	-	dull banded durain with vitrain bands
	0.02	-	intermediate, normal banded duroclarain
	0.04	-	dull, shaly durain with occasional vitrain micro-bands
	0.02	-	dull and bright banded clarodurain
	0.10	-	semi-bright clarain
SH81-4T			(-box 2)
	0.04	-	semi-bright clarain
	0.01	-	pyrite-rich (fusainous?) band
	0.03	-	intermediate duroclarain with occasional thin fusain bands
	0.01	-	dull fusain

SAMPLE PETROG./PALYN	THICKNESS(m)		LITHOTYPE/LITHOFACIES
	0.39	-	(lost core)
	0.02	-	semi-bright clarain
SH81-5T	0.02	-	intermediate to semi-bright clarain with occasional thin durain bands
	0.05	-	intermediate duroclarain to clarain with rare, thin fusain lenses
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	0.19	-	semi-bright clarain; pyrite on cleats
SH81-6T	0.14	-	intermediate to semi-bright clarain with minor amounts of duroclarain
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	0.03	-	intermediate duroclarain with 2mm fusain bands from 0.02-0.03m
	0.01	-	intermediate clarodurain
SH81-7T	0.06	-	semi-bright clarain; pyrite on cleats
	0.02	-	intermediate duroclarain
	0.09	-	semi-bright clarain with micro-bands of fusain

	0.03	-	intermediate duroclarain
	0.01	-	dull coaly shale
	0.02	-	intermediate duroclarain with 0.03mm fusain bands
SH81-8T	0.03	-	dull, micro-banded shaly durain

SAMPLE PETROG./PALYN	THICKNESS(m)		LITHOTYPE/LITHOFACIES
	0.03	-	semi-bright clarain with micro-bands of fusain
SH81-8T	0.06	-	semi-bright clarain
	0.01	-	dull coaly shale to shaly durain; fissile; bedding slickensides
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	0.03	-	intermediate duroclarain
		-	(26m marker)
	0.02	-	semi-bright clarain
SH81-9T	0.04	-	dull and bright banded clarodurain
	0.02	-	semi-bright duroclarain
	0.03	-	bright clarain
	0.03	-	semi-bright duroclarain
	0.02	-	semi-dull clarodurain
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	0.01	-	dull coaly shale
	0.03	-	dull micro-banded clarodurain with fusain lens
	0.03	-	dull, normal banded shaly coal (coaly shale and durain)
SH81-10T	0.01	-	dull coaly shale
	0.02	-	intermediate, micro-banded clarodurain
	0.04	-	dull black fusain
	0.07	-	semi-bright, micro-banded clarain, jointed and slickensided with 1mm-thick calcite and slickensided films on slicken's,

SAMPLE PETROG./PALYN	THICKNESS(m)		LITHOTYPE/LITHOFACIES
		-	thin fusain lenses at base
	0.02	-	intermediate duroclarain
	0.07	-	dull banded clarodurain with minor duroclarain
SH81-10T			
	0.02	-	intermediate, waxy, micro-banded clarodurain
	0.02	-	dull, micro-banded coaly shale
	0.04	-	bright, micro-banded duroclarain
	0.02	-	dull and bright, micro-banded clarain, interbanded with fusain
	0.08	-	semi-bright, normal banded clarain, grading downward to normal to micro-banded clarodurain from 0.03-0.08m
SH81-11T			
	0.005	-	dull fusain
	0.10	-	micro-banded clarodurain with 0.02m band of normal banded vitrain and fusain in center
	0.02	-	dull durain
	0.08	-	intermediate, micro-banded clarodurain to duroclarain
	0.07	-	semi-bright, micro-banded clarain with thin calcite film on cleats and irridescent blue film on joints and cleats
SH81-12T	0.04	-	intermediate duroclarain with calcite and blue hue

SAMPLE PETROG./PALYN	THICKNESS(m)		LITHOTYPE/LITHOFACIES
	0.02	-	dull durain and pyritic coaly shale with calcite and blue films
	0.06	-	intermediate to semi-bright, vitrainous, micro-banded clarain with thin fusain lenses, calcite veinlets and blueish film
SH81-12T	0.03	-	dull banded coaly shale and clarodurain with blueish hue and orange iron oxide coating on cleats
	0.02	-	semi-dull shaly coal (clarodurain)
	0.02	-	intermediate micro-banded clarodurain
	0.02	-	dull, microbanded coaly shale
	0.10		(- lost core)
	0.02	-	dull micro-banded shaly coal (durain with normal bands of vitrain)
	0.02	-	dull fusain
SH81-13T	0.05	-	dull normal and micro-banded durain with 1cm vitrain at top, fusain lenses below
	0.08	-	dull banded shaly coal (micro-banded coaly shale with normal-banded vitrain)
	0.02	-	dull shaly durain
	0.03	-	dull, micro-banded coaly shale
	0.10	-	(lost core)

------------------------ End of seam ------------------------

	0.03	-	dark grey to black, carbonaceous mudstone (Fc)

- (27.5m marker)

MACROSCOPIC PROFILE
SEAM: No. 3
LOCATION: Drillhole SH85 (PIEDMONT ZONE)

SAMPLE PETROG./PALYN	THICKNESS(m)		LITHOTYPE/LITHOFACIES
SH85-1T	0.06	-	bright coal impregnated with pyrite and calcite
	0.16	-	bright banded (clarain)
	0.02	-	dull banded
SH85-2T	0.10	-	grey mudstone (Fm)
	0.14	-	lost core
SH85-3T	0.13	-	dull and bright coal
	0.02	-	dull and bright coal
	0.03	-	shaly coal
	0.11	-	grey to black mudstone, coaly in places (Fc to Ci)
SH85-4T	0.05	-	black coaly shale
	0.14	-	dull and bright coal
SH85-5T	0.05	-	dull coal with minor fusain
	0.24	-	dull and bright, thin normal to micro-banded coal
	0.09	-	dull banded coal
SH85-6T	0.03	-	dull and bright coal
	0.10	-	dull banded coal
	0.13	-	dull and bright banded coal

SAMPLE PETROG./PALYN	THICKNESS(m)		LITHOTYPE/LITHOFACIES
	0.01	-	dull, fusainous coal
SH85-7T	0.17	-	dull and bright banded coal
	0.02	-	bright banded, calcite-impregnated coal
	0.02	-	dull banded, calcite-impregnated
	0.04	-	bright banded
	0.03	-	shaly coal
SH85-8T	0.03	-	black and dark grey, fissile coaly shale
	0.20	-	shaly coal with microbands of vitrain (<1mm)
	0.07	-	grey, carbonaceous mudstone
	0.03	-	black coaly shale
	0.03	-	shaly coal
	0.04	-	dull banded coal
SH85-9T	0.02	-	bright banded coal
	0.10	-	shaly, microbanded coal
	0.06	-	dull and bright micro-to normal banded coal (Bedding dips 12°)
	0.03	-	black, carbonaceous mudstone (Fc)
SH85-10T	0.05	-	shaly coal
		-	(14m drill marker)

SAMPLE PETROG./PALYN	THICKNESS(m)		LITHOTYPE/LITHOFACIES
	0.03	-	dull banded coal
	0.05	-	shaly coal, calcite veins infilling joints
SH85-10T	0.11	-	black coaly shale with calcite-infilled joints; 0.02m grey mudstone (Fm) lamina at 0.01m
	0.11	-	shaly coal
	0.06	-	grey and black laminated coaly shale
	0.17	-	shaly coal
			(- box 2)
SH85-11T	0.04	-	black, coaly shale
	0.05	-	bright banded coal
	0.06	-	dull and bright banded coal with pyrite lens at base
	0.07	-	pale grey, calcareous, fine-grained sandstone; coarsens upward (Sm)
SH85-12T	0.05	-	shaly, pyritic coal
	0.04	-	pale grey, calcareous, fine-grained sandstone (Sm)
SH85-13T	0.11	-	black to dark grey laminated, coaly mudstone with poorly preserved bark vitrain (<u>Calamites</u> <u>sp</u>.); Ci

SAMPLE PETROG./PALYN	THICKNESS(m)		LITHOTYPE/LITHOFACIES
SH85-14T	0.22	-	grey very fine-grained sandstone with rare thin bark vitrain bands (sp. indet.); sharp, irregular base. (Sm)
SH85-15T	0.57	-	grey, thinly to thickly laminated carbonaceous mudstone; (Fcl to Fc) thin laminae irregularly formed
		-	0.02m dull, waxy coaly shale at 0.07m capped by bark vitrain
		-	0.03m black shaly coal at 0.35m
SH85-16T	0.04	-	pyritic shaly coal with common 5mm-thick vitrain bands and bark vitrain (sp.indet.)
SH85-17T	0.12	-	dull, micro-banded durain (to clarodurain)
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	0.07	-	dull coaly shale with occasional vitrain bands
	0.01	-	grey load-cast mudstone
	0.05	-	muddy, interlaminated, loaded, calcite-impregnated coaly shale
SH85-18T			
	0.12	-	grey carbonaceous mudstone with rare vitrain bands at base
	0.06	-	coaly shale; occasional muddy laminae and pyrite lenses
	0.07	-	grey mudstone, with 1 cm thick dull coal and vitrain laminae

SAMPLE PETROG./PALYN	THICKNESS(m)		LITHOTYPE/LITHOFACIES
SH85-18T	0.05	-	coaly shale (Ci); thin mudstone (Fm) lamina at base
SH85-19T	0.14	-	shaly coal with common bark vitrain; calcite - impregnated basal 0.02m
SH85-20T	0.29	-	grey coarse siltstone with very thin slightly contorted laminae (Fl to Shl); rooted [- 17m marker]
SH85-21T	0.07	-	cm-scale interlaminated dull banded coal (clarodurain) and grey mudrock (Ci)
SH85-22T	0.29	-	grey siltstone with impure carbonaceous matter (Fcl) and rootlets; grading below 0.24 to Fc
SH85-23T	0.57	-	Ci; interlaminated carbonacous mudstone and dull waxy to microbanded (quasi-sapropelic) shaly coal to coaly shale; dulls upward overall. Base, gradational, dull, waxy, with 1mm thick pyrite lenses

SAMPLE PETROG./PALYN	THICKNESS(m)		LITHOTYPE/LITHOFACIES	
			Ci	Fc
			-	3 cm
			4	-
			-	4
			1	-
			-	1
			3	-
			-	0.5
			1	-
			-	3
			5	-
			-	1
			4	-
			-	3
			3	-
			-	2
SH85-24T	0.04	-	micro-banded, dull clarain to clarodurain with rare fusain clasts and three 0.5 - 1.0 cm - thick coaly shale laminae	
SH85-25T	0.02	-	shaly clarodurain with pyrite lenses at base	

SEAM: No. 7
LOCATION: Drillhole SH95 (INNER MIRE ZONE)
Roof: Sideritic mudrock (Fm) with numerous <u>Cordaites</u> compressions

SAMPLE	THICKNESS(m)		LITHOTYPE/LITHOFACIES
"	0.02m	-	microbanded semidull clarodurain
SH95-1V	0.02m	-	microbanded duroclarain
"	0.03m	-	microbanded, semi-bright clarain
SH95-2V	0.04m	-	microbanded durain
SH95-3V	0.09m	-	bright vitroclarain
SH95-4V	0.04m	-	semi-bright clarain to duroclarain
"	0.05m	-	microbanded, intermediate duroclarain
SH95-5V	0.03m	-	dull micro- to thinly banded shaly coal to coaly shale (Ci)
"	0.04m	-	semi-bright clarain
"	0.03m	-	intermediate duroclarain
"	0.02m	-	semi-dull, slightly shaly clarodurain
SH95-6V	0.01m	-	semi-bright clarain
"	0.03m	-	bright vitroclarain
"	0.03m	-	vitrain
SH95-7V	0.25m	-	semi-bright clarain
SH95-8V	0.09m	-	intermediate duroclarain (to clarodurain
SH95-9AV	0.03m	-	semi-bright clarain
"	0.04m	-	normal banded, intermediate duroclarain
SH95-9BV	0.02m	-	semi-bright clarain
"	0.03m	-	semi-bright to intermediate, thinly banded duroclarain

"	0.03m	-	semi-bright clarain
SAMPLE	THICKNESS(m)		LITHOTYPE/LITHOFACIES
SH95-10V	0.02m	-	laminated shaly coal with fusain and vitrain
"	0.04m	-	vitrain
"	0.15m	-	semi-bright clarain, micro-banded in places, grading in upper 0.04m to thickly banded vitroclarain with < 1mm thick fusain micro-bands
	Floor:	-	mudrock, Fm to Fc

Seam: 'Alpha'
Location: Drillhole SA.88.2
South Athol
Cumberland Basin
Depth: 3574.5' (1089.8m) at roof

Sample	Descripton	Thickness
Roof:	Dark grey, siderite-banded mudstone with common Cordaites foliage compressions.	
Alpha-1	Normal banded clarodurain (clarain with dull durain bands).	0.01m
Alpha-2	Coaly shale. 0.003m	
"	Bb: several mm-thick vitrain with thin microbanded clarodurain to shaly interbands in upper 0.02m; semi-bright vitroclarain below.	0.04m
Alpha-3	Interlaminated coaly shale and carbonaceous mudstone.	0.015m
Alpha-4	Microbanded clarain to duroclarain with very thin pyrite lens at top; grades below to:	0.015m
"	microbanded clarodurain.	0.02m
Alpha-5	Clarain to vitroclarain with common thin fusain lenses.	0.02m
Alpha-6	Clarodurain to duroclarain	0.015m
"	Dull, shaly(?) coal with vitrain bands (looks like clarodurain)	0.015m
Alpha-7	Bright vitroclarain with 0.01m vitrain near top.	0.04m
Alpha-8	Somewhate shaly clarodurain; grades to:	0.03m
Alpha-9	Dull/semi-dull durain with vitrain fragments; vitrain becomes banded at base.	0.015m
Alpha-10	Vitroclarain (core broken at base)	0.06m
Alpha-11	Slightly shaly duroclarain with occasional vitrain bands and fusain lenses.	0.025m
	Floor: rooted, several cm-scale interbedded very fine grained sandstone and darker grey silty mudstone.	

Seam: 'Epsilon'
Location: Drillhole SA.88.2
South Athol
Cumberland Basin
Depth: 3778.7' (1152.04m)

Sample	Description	Thickness
	Upper rider:	
E-1	Dark grey to black Fc	0.035m
"	Microbanded Ci (very shaly to coaly shale)	0.02m
"	Dark grey to black coaly shale to mudstone (Ci)	0.01m
"	Black coaly shale with 1mm-thick vitrain bands in upper 1cm.	0.06m
	Dark grey, internally slickensided, rooted, crudely laminated, carbonaceous mudstone; plant fossils very rare (Fcl).	0.50m
	Lower, main leaf:	
E-2	Microbanded shaly coal with 1mm-thick vitrain; 1cm layer of thick vitrain bands at 2.5cm.	0.06m
E-3	'Dirty' fusain.	0.015m
"	Pyritic & fusainous coaly shale.	0.01m
E-4	Dull black coaly shale with vitrain and possible fusain clasts (semi-cannel).	0.03m
E-5	Semi-dull to intermediate clarodurain (semi-cannel).	0.02m
E-6	Coaly shale with occasional vitrain microbands, grading to carbonaceous mudstone in places and at top to clarodurain.	0.07m
E-7	Dull and bright banded coal: 0.05 to 1cm thick clarodurain, 1 to 3mm + vitrain and <0.05cm black microbanded carbonaceous shale, with occasional pyrite lenses and laminae.	0.09m
E-8	Same as above, minus thicker laminae of coaly black mudstone (shale; brightens upward from base of seam. (vitrain-banded clarodurain to duroclarain, rare fusain laminae).	0.09m

Sample	Description	Thickness
E-9	Thinly banded semi-dull clarodurain with vitrain; somewhat shaly.	0.07m
E-10	Black, microbanded coaly shale, grading upward to less shaly clarodurain; base of seam gradational.	
	Floor: rooted, carbonaceous, very fine grained sandstone to silty sandstone.	

Table C1. Maceral and mineral composition (Volume %) No. 3 Seam, SH 81

	VITRINITE							LIPTINITE									INERTINITE							
SAMPLE ID	Ps1	Ps2	Tc	Gc	Cc	Dc	TOTAL	Sp	Cut	Sub	Res	Ld	Al	Am	Ex	TOTAL	F	Sf	Ma	Sc	Id	Mi	TOTAL	MM
	6.5	2.0	43.3	13.1	5.7	6.9	77.6	2.4	1.2	0.4	0.8	0.8	1.2	0.0	0.4	7.3	4.9	2.9	1.8	0.2	4.1	1.2	15.1	
SH-81-1T	6.4	2.0	42.4	12.8	5.6	6.8	76.0	2.4	1.2	0.4	0.8	0.8	1.2	0.0	0.4	7.2	4.8	2.8	1.8	0.2	4.0	1.2	14.8	2.0
	3.4	1.1	33.1	6.2	9.8	7.7	61.3	5.6	0.6	0.2	2.4	3.2	1.5	0.0	0.6	14.1	7.1	7.3	3.2	0.2	5.8	1.1	24.6	
SH-81-2T	3.2	1.0	31.0	5.8	9.2	7.2	57.4	5.2	0.6	0.2	2.2	3.0	1.4	0.0	0.6	13.2	6.6	6.8	3.0	0.2	5.4	1.0	23.0	6.4
	4.5	2.1	27.9	12.4	11.6	6.6	65.1	6.2	0.6	0.0	2.7	1.0	3.3	0.0	0.0	13.8	4.1	8.7	1.4	0.4	6.2	0.2	21.1	
SH-81-3T	4.4	2.0	27.0	12.0	11.2	6.4	63.0	6.0	0.6	0.0	2.6	1.0	3.2	0.0	0.0	13.4	4.0	8.4	1.4	0.4	6.0	0.2	20.4	3.2
	3.2	3.8	24.0	18.3	6.1	4.0	59.4	4.0	1.1	0.0	3.2	2.5	2.5	0.0	0.6	13.9	7.8	6.3	2.9	0.8	7.8	1.1	26.7	
SH-81-4T	3.0	3.6	22.8	17.4	5.8	3.8	56.4	3.8	1.0	0.0	3.0	2.4	2.4	0.0	0.6	13.2	7.4	6.0	2.8	0.8	7.4	1.0	25.4	5.0
	9.6	1.5	25.6	23.3	10.8	4.7	75.5	3.9	0.2	0.0	2.2	2.4	1.4	0.0	0.0	10.1	4.6	5.1	0.6	0.2	3.2	0.6	14.5	
SH-81-5T	9.2	1.4	24.4	22.2	10.3	4.5	72.0	3.7	0.2	0.0	2.1	2.3	1.3	0.0	0.0	9.6	4.4	4.9	0.6	0.2	3.1	0.6	13.8	4.6
	9.9	6.1	25.5	26.7	9.9	2.2	80.4	2.8	0.2	0.4	2.0	0.8	1.2	0.0	0.2	7.7	3.6	3.8	0.8	0.2	2.6	0.8	11.9	
SH-81-6T	9.8	6.0	25.2	26.4	9.8	2.2	79.4	2.8	0.2	0.4	2.0	0.8	1.2	0.0	0.2	7.6	3.6	3.8	0.8	0.2	2.6	0.8	11.8	1.2
	4.0	5.9	21.9	38.1	5.7	2.4	77.9	2.4	0.2	0.2	2.2	1.0	1.0	0.0	0.2	7.3	4.5	5.7	0.6	0.2	3.4	0.4	14.8	
SH-81-7T	4.0	5.8	21.6	37.6	5.6	2.4	77.0	2.4	0.2	0.2	2.2	1.0	1.0	0.0	0.2	7.2	4.4	5.6	0.6	0.2	3.4	0.4	14.6	1.2
	3.9	7.7	17.0	27.3	5.6	1.7	63.1	4.6	0.4	0.2	2.7	1.9	2.3	0.0	0.4	12.4	7.7	7.2	2.1	0.4	6.4	0.6	24.4	
SH-81-8T	3.8	7.4	16.4	26.4	5.4	1.6	61.0	4.4	0.4	0.2	2.6	1.8	2.2	0.0	0.4	12.0	7.4	7.0	2.0	0.4	6.2	0.6	23.6	3.4
	1.7	6.5	17.5	31.4	8.8	1.5	67.4	6.9	1.5	0.0	0.6	0.4	4.2	0.0	0.0	13.7	5.5	5.7	1.5	0.4	5.7	0.2	18.9	
SH-81-9T	1.6	6.2	16.6	29.8	8.4	1.4	64.0	6.6	1.4	0.0	0.6	0.4	4.0	0.0	0.0	13.0	5.2	5.4	1.4	0.4	5.4	0.2	18.0	5.0
	2.8	6.0	10.9	27.6	9.4	2.1	58.8	4.9	0.6	0.2	1.3	1.5	2.8	0.2	0.2	11.8	11.5	6.4	0.6	0.2	9.8	0.9	29.5	
SH-81-10T	2.6	5.6	10.2	25.8	8.8	2.0	55.0	4.6	0.6	0.2	1.2	1.4	2.6	0.2	0.2	11.0	10.8	6.0	0.6	0.2	9.2	0.8	27.6	6.4
	4.6	11.3	16.9	17.9	12.5	1.0	64.2	4.8	1.7	0.2	1.7	1.3	2.5	1.3	0.6	14.0	7.5	6.0	0.6	0.0	7.1	0.6	21.9	
SH-81-11T	4.4	10.8	16.2	17.2	12.0	1.0	61.6	4.6	1.6	0.2	1.6	1.2	2.4	1.2	0.6	13.4	7.2	5.8	0.6	0.0	6.8	0.6	21.0	4.0
	3.9	8.4	19.1	24.0	14.8	7.1	77.1	6.4	0.6	0.0	0.9	3.9	1.7	0.2	0.6	14.3	0.6	5.6	0.0	0.0	2.4	0.0	8.6	
SH-81-12T	3.6	7.8	17.8	22.4	13.8	6.6	72.0	6.0	0.6	0.0	0.8	3.6	1.6	0.2	0.6	13.4	0.6	5.2	0.0	0.0	2.2	0.0	8.0	6.6
	2.4	6.9	16.3	16.7	3.6	1.0	46.9	11.7	3.6	0.0	1.9	1.2	7.2	1.0	0.5	27.0	12.7	6.2	1.0	0.0	6.0	0.2	26.1	
SH-81-13T	2.0	5.8	13.6	14.0	3.0	0.8	39.2	9.8	3.0	0.0	1.6	1.0	6.0	0.8	0.4	22.6	10.6	5.2	0.8	0.0	5.0	0.2	21.8	16.4

VITRINITE:
- Ps1 Pseudovitrinite derived from telocollinite
- Ps2 Pseudovitrinite derived from gelocollinite
- Tc Telocollinite/Telinite
- Gc Gelocollinite
- Cc Corpocollinite
- Dc Desmocollinite

LIPTINITE:
- Sp Sporinite
- Cut Cutinite
- Sub Suberinite
- Res Resinite/Fluorinite
- Ld Liptodetrinite
- Al Alginite
- Am Amorphous Liptinite
- Ex Exsudatinite

INERTINITE:
- F Fusinite
- Sf Semifusinite
- Ma Macrinite
- Sc Sclerotinite
- Id Inertodetrinite
- Mi Micrinite

Note: The shaded values are mineral matter free.

Table C2. Maceral and mineral composition (Volume %) No. 3 Seam, SH 85

	VITRINITE							LIPTINITE									INERTINITE							
SAMPLE ID	Ps1	Ps2	Tc	Gc	Cc	Dc	TOTAL	Sp	Cut	Sub	Res	Ld	Al	Am	Ex	TOTAL	F	Sf	Ma	Sc	Id	Mi	TOTAL	MM
	2.8	4.0	19.9	27.2	28.3	4.2	86.4	1.9	0.7	0.0	2.3	0.9	1.6	0.0	1.6	9.1	1.2	0.7	1.2	0.0	0.5	0.5	4.0	
SH-85-1T	2.4	3.4	17.0	23.2	24.2	3.6	73.8	1.6	0.6	0.0	2.0	0.8	1.4	0.0	1.4	7.8	1.0	0.6	1.0	0.0	0.4	0.4	3.4	14.6
	0.0	0.0	26.1	43.5	0.0	0.0	69.6	4.3	0.0	0.0	0.0	4.3	0.0	0.0	0.0	8.7	17.4	0.9	0.0	0.0	3.5	0.0	21.7	
SH-85-2T	0.0	0.0	6.0	10.0	0.0	0.0	16.0	1.0	0.0	0.0	0.0	1.0	0.0	0.0	0.0	2.0	4.0	0.2	0.0	0.0	0.8	0.0	5.0	77.0
	2.9	5.4	5.6	18.6	4.3	1.1	37.9	8.3	1.1	0.0	2.0	1.8	2.0	2.0	0.7	17.9	27.1	5.4	4.5	0.4	5.4	1.3	44.2	
SH-85-3T	2.6	4.8	5.0	16.6	3.8	1.0	33.8	7.4	1.0	0.0	1.8	1.6	1.8	1.8	0.6	16.0	24.2	4.8	4.0	0.4	4.8	1.2	39.4	10.8
	0.9	0.9	8.2	27.3	6.5	0.9	44.6	10.0	0.4	0.0	3.5	2.6	2.2	0.0	0.4	19.0	13.4	3.5	3.5	0.4	12.1	3.5	36.4	
SH-85-4T	0.4	0.4	3.8	12.6	3.0	0.4	20.6	4.6	0.2	0.0	1.6	1.2	1.0	0.0	0.2	8.8	6.2	1.6	1.6	0.2	5.6	1.6	16.8	53.8
	5.6	5.8	17.7	24.9	6.4	1.9	62.4	10.6	1.0	0.2	3.1	1.5	1.5	0.0	0.8	18.7	8.5	3.5	2.9	0.0	2.9	1.0	18.9	
SH-85-5T	5.4	5.6	17.0	24.0	6.2	1.8	60.0	10.2	1.0	0.2	3.0	1.4	1.4	0.0	0.8	18.0	8.2	3.4	2.8	0.0	2.8	1.0	18.2	3.8
	4.2	3.7	10.6	25.6	11.8	2.3	58.2	10.2	3.7	0.0	1.4	1.2	3.0	0.0	2.3	21.7	14.1	2.3	0.5	0.0	2.3	0.9	20.1	
SH-85-6T	3.6	3.2	9.2	22.2	10.2	2.0	50.4	8.8	3.2	0.0	1.2	1.0	2.6	0.0	2.0	18.8	12.2	2.0	0.4	0.0	2.0	0.8	17.4	13.4
	3.7	4.2	21.1	37.3	4.2	1.0	71.6	11.4	1.0	0.0	2.5	0.2	1.2	0.0	1.0	17.4	4.7	3.2	0.5	0.0	1.5	1.0	10.9	
SH-85-7T	3.0	3.4	17.0	30.0	3.4	0.8	57.6	9.2	0.8	0.0	2.0	0.2	1.0	0.0	0.8	14.0	3.8	2.6	0.4	0.0	1.2	0.8	8.8	19.6
	0.3	1.5	12.0	29.1	5.1	1.5	49.5	15.9	1.5	0.3	2.1	2.1	1.2	0.0	1.2	24.3	21.6	2.1	0.0	0.0	0.9	1.5	26.1	
SH-85-8T	0.2	1.0	8.0	19.4	3.4	1.0	33.0	10.6	1.0	0.2	1.4	1.4	0.8	0.0	0.8	16.2	14.4	1.4	0.0	0.0	0.6	1.0	17.4	33.4
	4.7	3.1	14.5	30.0	7.2	2.0	61.5	11.0	1.8	0.0	1.3	2.7	4.9	0.0	0.4	22.1	8.1	2.5	0.4	0.7	3.1	1.6	16.3	
SH-85-9T	4.2	2.8	13.0	26.8	6.4	1.8	55.0	9.8	1.6	0.0	1.2	2.4	4.4	0.0	0.4	19.8	7.2	2.2	0.4	0.6	2.8	1.4	14.6	10.6
	1.7	0.8	22.7	21.8	6.4	1.1	54.4	12.7	1.4	0.0	2.2	2.5	4.7	0.6	0.6	24.6	9.4	6.6	2.8	0.3	0.8	1.1	21.0	
SH-85-10T	1.2	0.6	16.4	15.8	4.6	0.8	39.4	9.2	1.0	0.0	1.6	1.8	3.4	0.4	0.4	17.8	6.8	4.8	2.0	0.2	0.6	0.8	15.2	27.6
	3.3	1.4	22.2	25.1	8.5	2.0	62.7	13.6	1.0	0.0	1.7	1.3	3.3	0.0	0.0	20.9	10.9	2.4	1.7	0.0	1.4	0.0	16.4	
SH-85-11T	2.8	1.2	18.6	21.0	7.1	1.7	52.4	11.4	0.8	0.0	1.4	1.1	2.8	0.0	0.0	17.5	9.1	2.0	1.4	0.0	1.2	0.0	13.7	16.4
	3.6	4.6	19.1	31.3	1.0	2.6	62.2	13.2	1.3	0.0	4.9	1.6	3.9	0.0	1.0	26.0	4.9	4.9	1.6	0.3	0.0	0.0	11.8	
SH-85-12T/13T	2.2	2.8	11.6	19.0	0.6	1.6	37.8	8.0	0.8	0.0	3.0	1.0	2.4	0.0	0.6	15.8	3.0	3.0	1.0	0.2	0.0	0.0	7.2	39.2
	0.0	3.3	9.6	28.2	2.6	0.5	44.2	9.1	1.0	0.0	4.8	0.0	6.3	0.0	1.5	22.6	24.6	7.6	1.0	0.0	0.0	0.0	33.2	
SH-85-15T/16T	0.0	2.0	5.8	17.1	1.6	0.3	26.8	5.5	0.6	0.0	2.9	0.0	3.8	0.0	0.9	13.7	14.9	4.6	0.6	0.0	0.0	0.0	20.1	39.4
	1.4	8.0	14.2	29.0	8.9	2.7	64.2	12.6	1.1	0.0	1.8	1.6	3.4	0.0	0.5	21.0	5.9	5.7	0.9	0.2	1.1	0.9	14.8	
SH-85-17T	1.2	7.0	12.4	25.4	7.8	2.4	56.2	11.0	1.0	0.0	1.6	1.4	3.0	0.0	0.4	18.4	5.2	5.0	0.8	0.2	1.0	0.8	13.0	12.4
	4.4	3.7	14.5	31.0	4.0	0.7	58.2	17.2	1.0	0.0	2.4	2.4	4.0	0.0	1.3	28.3	9.4	1.7	0.3	0.0	1.7	0.3	13.5	
SH-85-18T	2.6	2.2	8.6	18.4	2.4	0.4	34.6	10.2	0.6	0.0	1.4	1.4	2.4	0.0	0.8	16.8	5.6	1.0	0.2	0.0	1.0	0.2	8.0	40.6
	0.5	8.1	17.6	43.5	5.3	0.8	75.8	10.4	1.0	0.0	1.5	2.0	0.8	0.0	0.0	15.8	4.6	1.5	1.3	0.3	0.5	0.3	8.4	
SH-85-19T	0.4	6.4	13.8	34.2	4.2	0.6	59.6	8.2	0.8	0.0	1.2	1.6	0.6	0.0	0.0	12.4	3.6	1.2	1.0	0.2	0.4	0.2	6.6	21.4
	0.0	0.0	13.2	31.4	4.6	0.4	49.6	23.6	2.0	0.0	1.6	2.8	6.2	0.0	5.0	41.2	6.0	0.8	0.4	0.4	1.2	0.4	9.2	
SH-85-21T	0.0	0.0	6.6	15.7	2.3	0.2	24.8	11.8	1.0	0.0	0.8	1.4	3.1	0.0	2.5	20.6	3.0	0.4	0.2	0.2	0.6	0.2	4.6	50.0
	0.0	1.0	19.9	26.4	7.4	1.0	55.6	23.8	1.6	0.0	1.3	1.3	4.8	0.0	2.3	35.0	3.9	3.9	1.0	0.0	0.6	0.0	9.3	
SH-85-23T/25T	0.0	0.6	12.4	16.4	4.6	0.6	34.6	14.8	1.0	0.0	0.8	0.8	3.0	0.0	1.4	21.8	2.4	2.4	0.6	0.0	0.4	0.0	5.8	37.8

VITRINITE :
- Dc Desmocollinite
- Ps1 Pseudovitrinite derived from telocollinite
- Ps2 Pseudovitrinite derived from gelocollinite
- Tc Telocollinite/Telinite
- Gc Gelocolinite
- Cc Corpocollinite

LIPTINITE:
- Sp Sporinite
- Cut Cutinite
- Sub Suberinite
- Res Resinite/Fluorinite
- Ld Liptodetrinite
- Al Alginite
- Am Amorphous Liptinite
- Ex Exsudatinite

INERTINITE :
- F Fusinite
- Sf Semifusinite
- Ma Macrinite
- Sc Sclerotinite
- Id Inertodetrinite
- Mi Micrinite

Note: Shaded values are mineral matter free.

Table C3. Maceral and mineral composition (Volume %) No. 3 Seam, SH 72

	VITRINITE							LIPTINITE									INERTINITE							
SAMPLE ID	Ps1	Ps2	Tc	Gc	Cc	Dc	TOTAL	Sp	Cut	Sub	Res	Ld	Al	Am	Ex	TOTAL	F	Sf	Ma	Sc	Id	Mi	TOTAL	MM
	1.1	0.6	8.6	9.7	3.4	0.0	23.4	14.3	1.7	0.0	1.1	1.7	1.7	0.0	1.7	22.3	21.7	7.4	7.4	0.6	11.4	5.7	54.3	
SH-72-1UT/2UT	0.4	0.2	3.0	3.4	1.2	0.0	8.2	5.0	0.6	0.0	0.4	0.6	0.6	0.0	0.6	7.8	7.6	2.6	2.6	0.2	4.0	2.0	19.0	65.0
	1.9	2.1	15.2	20.6	9.8	0.5	50.1	9.1	2.8	0.0	1.6	2.1	2.1	0.0	0.7	18.5	11.0	4.0	5.2	0.5	10.1	0.7	31.4	
SH-72-3UT	1.6	1.8	13.0	17.6	8.4	0.4	42.8	7.8	2.4	0.0	1.4	1.8	1.8	0.0	0.6	15.8	9.4	3.4	4.4	0.4	8.6	0.6	26.8	14.6
	0.2	0.7	18.3	22.5	5.6	0.0	47.3	12.2	2.6	0.0	2.6	0.7	2.8	0.0	0.0	20.8	11.0	4.9	7.5	0.5	7.5	0.5	31.9	
SH-72-5UT	0.2	0.6	15.6	19.2	4.8	0.0	40.4	10.4	2.2	0.0	2.2	0.6	2.4	0.0	0.0	17.8	9.4	4.2	6.4	0.4	6.4	0.4	27.2	14.6
	7.1	2.1	25.4	29.6	9.2	1.3	74.6	6.5	1.0	0.2	2.5	3.1	1.7	0.0	0.6	15.7	3.8	2.7	0.2	0.0	2.7	0.2	9.6	
SH-72-6UT	6.8	2.0	24.2	28.2	8.8	1.2	71.2	6.2	1.0	0.2	2.4	3.0	1.6	0.0	0.6	15.0	3.6	2.6	0.2	0.0	2.6	0.2	9.2	4.6
	0.0	0.0	27.9	27.7	7.5	1.1	64.2	5.4	2.4	0.0	0.6	0.2	0.9	0.0	1.3	10.7	9.2	6.0	2.4	0.2	7.1	0.2	25.1	
SH-72-7UT	0.0	0.0	26.0	25.8	7.0	1.0	59.8	5.0	2.2	0.0	0.6	0.2	0.8	0.0	1.2	10.0	8.6	5.6	2.2	0.2	6.6	0.2	23.4	6.8
	2.7	1.6	19.0	35.8	12.9	2.9	74.8	6.7	1.6	0.4	1.0	1.2	0.6	0.0	0.8	12.5	3.7	6.3	0.8	0.0	1.4	0.4	12.7	
SH-72-8UT	2.6	1.6	18.6	35.0	12.6	2.8	73.2	6.6	1.6	0.4	1.0	1.2	0.6	0.0	0.8	12.2	3.6	6.2	0.8	0.0	1.4	0.4	12.4	2.2
	0.4	0.4	19.0	35.2	7.6	1.3	63.9	4.9	2.7	0.4	3.8	2.1	2.1	0.0	1.1	17.1	5.7	3.4	3.4	0.4	5.1	1.1	19.0	
SH-72-9UT	0.4	0.4	18.0	33.4	7.2	1.2	60.6	4.6	2.6	0.4	3.6	2.0	2.0	0.0	1.0	16.2	5.4	3.2	3.2	0.4	4.8	1.0	18.0	5.2
	0.8	0.4	22.6	44.1	8.1	2.1	78.1	6.2	0.4	0.0	1.7	2.3	0.4	0.0	0.8	11.8	5.4	2.1	1.0	0.0	1.4	0.2	10.1	
SH-72-10UT	0.8	0.4	21.8	42.6	7.8	2.0	75.4	6.0	0.4	0.0	1.6	2.2	0.4	0.0	0.8	11.4	5.2	2.0	1.0	0.0	1.4	0.2	9.8	3.4
	0.5	3.7	6.7	33.0	8.1	0.7	52.7	12.2	6.9	0.2	3.0	1.2	3.0	0.0	0.5	27.0	9.0	5.1	3.7	0.0	2.3	0.2	20.3	
SH-72-11UT	0.4	3.2	5.8	28.6	7.0	0.6	45.6	10.6	6.0	0.2	2.6	1.0	2.6	0.0	0.4	23.4	7.8	4.4	3.2	0.0	2.0	0.2	17.6	13.4
	1.7	0.9	14.2	43.0	2.0	0.0	61.8	10.5	11.4	0.0	2.3	4.6	6.6	0.0	1.4	36.8	0.3	1.1	0.3	0.6	0.0	0.3	2.6	
SH-72-1LT	1.2	0.6	10.0	30.2	1.4	0.0	43.4	7.4	8.0	0.0	1.6	3.2	4.6	0.0	1.0	25.8	0.2	0.8	0.2	0.4	0.0	0.2	1.8	29.8
	0.6	0.0	8.0	48.1	6.0	0.6	63.2	14.0	6.3	0.0	1.4	5.7	1.4	0.0	2.0	30.8	3.7	0.6	0.6	0.3	0.6	0.3	6.0	
SH-72-3LT	0.4	0.0	5.6	33.8	4.2	0.4	44.4	9.8	4.4	0.0	1.0	4.0	1.0	0.0	1.4	21.6	2.6	0.4	0.4	0.2	0.4	0.2	4.2	29.8
	3.4	0.0	7.7	27.7	3.4	0.0	42.1	23.0	1.7	0.0	1.3	3.8	3.0	0.0	3.4	36.2	14.0	2.1	1.7	0.0	2.1	1.7	21.7	
SH-72-5LT	1.6	0.0	3.6	13.0	1.6	0.0	19.8	10.8	0.8	0.0	0.6	1.8	1.4	0.0	1.6	17.0	6.6	1.0	0.8	0.0	1.0	0.8	10.2	53.0
	0.0	0.5	13.3	50.8	8.8	0.0	73.4	8.8	1.4	0.0	3.3	2.9	1.4	0.0	1.9	19.7	3.1	0.5	2.9	0.0	0.5	0.0	6.9	
SH-72-6LT	0.0	0.4	11.2	42.8	7.4	0.0	61.8	7.4	1.2	0.0	2.8	2.4	1.2	0.0	1.6	16.6	2.6	0.4	2.4	0.0	0.4	0.0	5.8	15.8
	0.0	0.0	7.8	45.6	10.4	0.0	63.8	22.1	3.3	0.0	2.0	1.6	1.6	0.0	1.3	31.9	2.0	0.3	1.3	0.0	0.3	0.3	4.2	
SH-72-7LT/8LT	0.0	0.0	4.8	28.0	6.4	0.0	39.2	13.6	2.0	0.0	1.2	1.0	1.0	0.0	0.8	19.6	1.2	0.2	0.8	0.0	0.2	0.2	2.6	38.6
	0.7	0.2	7.5	48.1	4.6	0.5	61.7	17.7	2.4	0.0	1.7	2.2	2.2	0.0	1.0	27.2	5.6	1.2	2.9	0.0	1.0	0.5	11.2	
SH-72-10LT/11LT	0.6	0.2	6.2	39.6	3.8	0.4	50.8	14.6	2.0	0.0	1.4	1.8	1.8	0.0	0.8	22.4	4.6	1.0	2.4	0.0	0.8	0.4	9.2	17.6

VITRINITE:
- Ps1 Pseudovitrinite derived from telocollinite
- Ps2 Pseudovitrinite derived from gelocollinite
- Tc Telocollinite/Telinite
- Gc Gelocollinite
- Cc Corpocollinite
- Dc Desmocollinite

LIPTINITE:
- Sp Sporinite
- Cut Cutinite
- Sub Suberinite
- Res Resinite/Fluorinite
- Ld Liptodetrinite
- Al Alginite
- Am Amorphous Liptinite
- Ex Exsudatinite

INERTINITE:
- F Fusinite
- Sf Semifusinite
- Ma Macrinite
- Sc Sclerotinite
- Id Inertodetrinite
- Mi Micrinite

Note: Shaded values are mineral matter free.

MAJOR OXIDE	1 SH-70-1UT-3UT	2 SH-70-1LT-4LT	3 SH-72-1UT-4UT	4 SH-72-5UT-11UT	5 SH-72-1LT-13LT	6 SH-81-1T-13T	7 SH-85-1T-13T	8 SH-85-17T-19T	9 SH-85-23T-25T
SiO2	45.0	41.8	56.2	48.2	51.6	51.7	45.4	44.7	52.1
Al2O3	18.9	16.0	24.3	16.1	22.0	21.3	17.9	17.7	20.8
CaO	6.44	5.66	0.27	1.94	0.68	2.77	7.20	9.17	1.24
MgO	1.38	1.02	1.65	0.59	1.56	1.23	2.23	2.63	2.87
Na2O	0.17	0.24	0.15	0.29	0.17	0.14	0.89	1.12	1.49
K2O	1.21	1.83	4.77	1.36	4.04	1.54	1.31	1.42	1.73
Fe2O3	17.7	24.3	7.61	26.1	16.1	15.4	17.0	14.3	13.0
MnO	0.12	0.08	0.0	0.01	0.03	0.03	0.20	0.26	0.10
TiO2	1.13	0.94	0.87	1.1	0.84	1.33	1.12	1.06	1.28
P2O5	0.06	0.10	0.04	0.02	0.09	0.14	0.15	0.15	0.19
LOI	2.54	4.16	3.93	2.85	3.16	2.85	4.62	6.00	3.16
TOTAL	94.6	96.1	99.8	98.6	100.3	98.4	98.0	98.5	98.0

* 1 : 1UT;2UT;3UT
2 : 1LT;3LT;4LT
3 : 1+2UT;3UT;4UT
4 : 5UT;6UT;7UT;8UT;9UT;10UT;11UT
5 : 1LT;3LT;5LT;6LT;7LT;8LT;10LT;11LT;13LT
6 : 1T;2T;3T;4T;5T;6T;7T;8T;9T;10T;11T;12T;13T
7 : 1T;3T;5T;6T;7T;8T;9T;10T;11T;12T;13T
8 : 17T;18T;19T
9 : 23T;24T;25T

Table C-4 Major oxide results from composite samples* of the No. 3 seam.

ELEMENT	SH-70-1UT-3UT	SH-70-1LT-4LT	SH-72-1UT-4UT	SH-72-5UT-11UT	SH-72-1LT-13LT	SH-81-1T-13T	SH-85-1T-13T	SH-85-17T-19T	SH-85-23T-25T
Ag ppm	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5
As ppm	210	210	22.0	590	210	350	130	130	100
Au ppb	<26	700	4	<10	<28	---	12	<10	4
B ppm	370	190	160	140	330	190	50	20	40
Ba ppm	286	453	574	404	520	470	469	342	501
Be ppm	15	36	5	32	10	11	8	8	6
Bi ppm	0.3	0.4	0.4	0.5	0.5	0.4	0.3	0.2	0.2
Cd ppm	1.0	2.4	2.0	3.8	4.4	0.6	5.8	7.4	3.0
Co ppm	36	31	16	52	36	39	43	35	36
Cr ppm	110	130	110	77	133	120	140	140	120
Cs ppm	10	12	15	12	20	11	7	5	6
Cu ppm	100	150	40.0	140	98.0	180	160	140	140
Ga ppm	17.1	16.7	26.4	18.7	25.2	18.9	16.3	16.9	19.4
Ge ppm	<10	<10	10	50	<10	10	<10	<10	<10
Hf ppm	2	4	4	3	3	3	2	2	2
Hg ppb	---	750	11	400	1000	31	72	45	14
In ppm	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5
Li ppm	120	86	110	60	130	140	76	82	120
Mo ppm	43	94	6	48	28	109	59	36	32
Nb ppm	10	13	12	12	11	12	9	7	10
Ni ppm	88	140	54	130	93	85	74	64	69
Pb ppm	260	400	52	660	780	380	460	260	280
Rb ppm	76	113	239	95	217	84	59	60	70
Sb ppm	6.8	26.0	2.9	19.0	12.0	3.9	3.4	2.0	2.5
Sc ppm	21	26	25	24	26	25	30	20	27
Se ppm	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5	<0.5
Sn ppm	17	13	4	31	8	32	16	22	44
Sr ppm	187	442	177	321	150	413	116	91	47
Ta ppm	2	2	2	2	2	2	1	1	1
Th ppm	7.9	9.8	14.0	10.5	15.0	10.1	4.9	4.2	4.7
Tl ppm	2.6	4.1	2.1	4.0	3.2	5.4	2.6	1.5	1.4
U ppm	2.2	3.5	4.4	4.0	5.9	2.5	1.5	1.4	1.4
V ppm	470	230	170	150	230	650	1200	1100	720
W ppm	3	5	3	5	3	3	2	2	2
Y ppm	42	109	16	90	31	44	35	36	20
Zn ppm	840	1400	970	1400	1200	650	2200	1600	850
Zr ppm	218	221	123	314	152	312	126	130	177
La ppm	23.3	24.3	37.0	18.0	43.6	33.5	17.9	18.4	15.3
Ce ppm	50.7	65.9	78.2	42.6	85.6	76.2	43.6	41.2	37.6
Pr ppm	6.6	9.7	9.4	6.1	10.3	9.5	5.4	5.4	4.8
Nd ppm	29.4	50.6	38.0	31.0	43.1	41.9	25.1	24.1	21.7
Sm ppm	7.2	14.7	5.6	9.5	8.3	8.7	5.4	5.3	4.2
Eu ppm	1.88	4.58	1.13	2.68	1.96	2.09	1.53	1.52	1.12
Gd ppm	9.2	22.3	3.9	14.5	7.3	9.4	6.4	6.2	4.5
Tb ppm	1.3	3.3	0.5	2.4	1.1	1.3	1.0	0.9	0.6
Dy ppm	7.1	18.5	3.4	14.1	6.8	7.6	5.8	6.0	4.1
Ho ppm	1.40	3.78	0.78	2.88	1.37	1.50	1.29	1.33	0.87
Er ppm	3.4	8.9	2.1	8.0	3.3	4.2	3.3	3.6	2.4
Tm ppm	0.4	1.0	0.2	0.9	0.4	0.5	0.4	0.4	0.2
Yb ppm	2.8	7.0	2.3	6.2	3.4	3.5	2.9	3.2	2.1
Lu ppm	0.49	0.99	0.41	0.88	0.52	0.51	0.49	0.50	0.37

Table C-5 Trace element results (in ash) from composite samples of the No. 3 seam.

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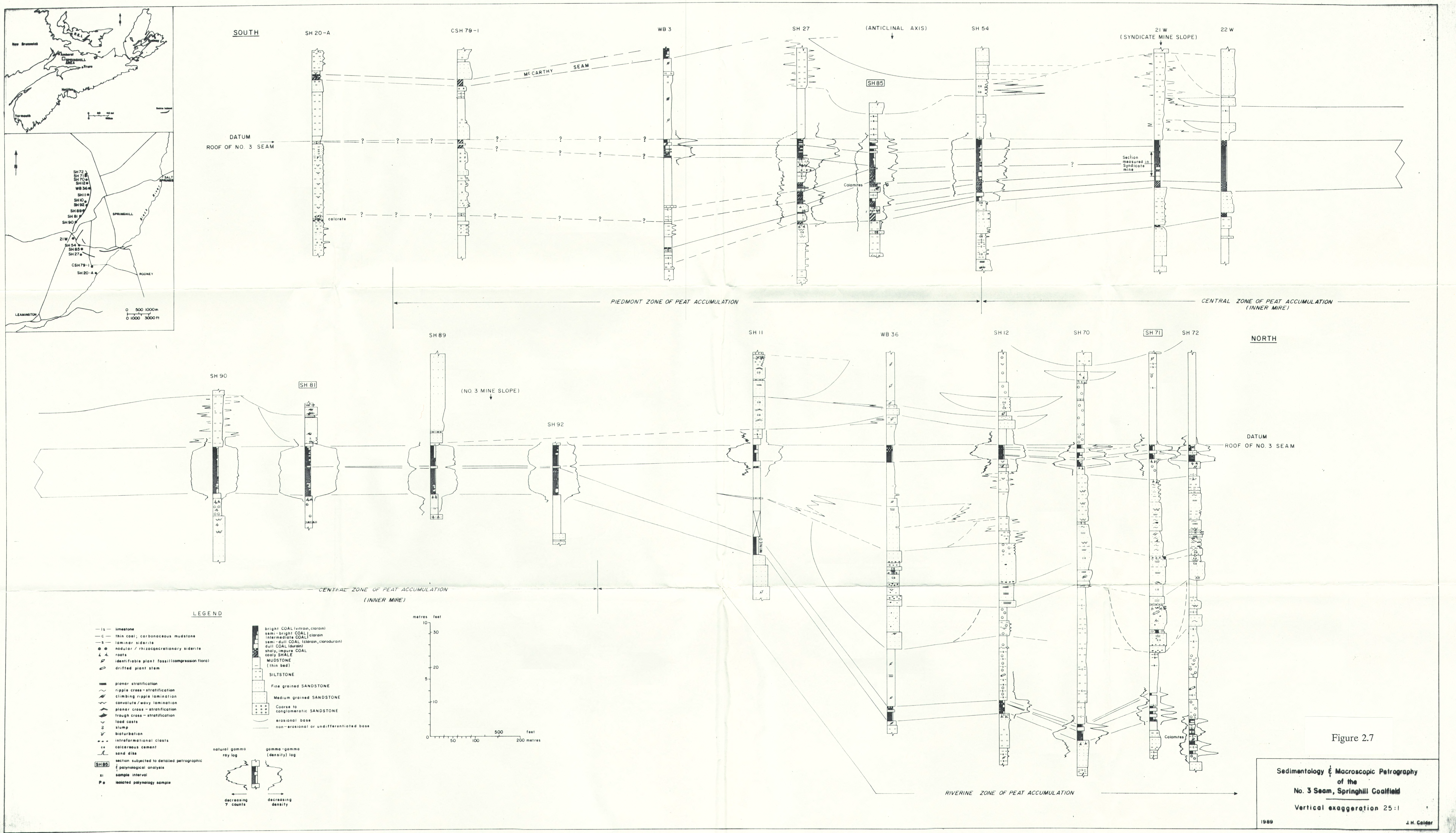


SOUTH
SH 20-A
CSH 79-1
WB 3
SH 27
(ANTICLINAL AXIS)
SH 54
21 W
(SYNDICATE MINE SLOPE)
22 W
McCARTHY SEAM
SH 85
DATUM
ROOF OF NO. 3 SEAM
calcrete
Calamites
Section measured in Syndicate mine
PIEDMONT ZONE OF PEAT ACCUMULATION
CENTRAL ZONE OF PEAT ACCUMULATION
(INNER MIRE)
SH 90
SH 81
SH 89
(NO. 3 MINE SLOPE)
SH 92
SH 11
MINED
WB 36
SH 12
SH 70
SH 71
SH 72
NORTH
DATUM
ROOF OF NO. 3 SEAM
Calamites
CENTRAL ZONE OF PEAT ACCUMULATION
(INNER MIRE)
RIVERINE ZONE OF PEAT ACCUMULATION
New Brunswick
P.E.I.
SPRINGHILL AREA
Truro
Halifax
Yarmouth
Sable Island
SALT SPRINGS
SPRINGHILL
ROONEY
LEAMINGTON
0 500 1000m
0 1000 3000 ft
LEGEND
ls limestone
c thin coal; carbonaceous mudstone
s laminar siderite
nodular / rhizoconcretionary siderite
roots
identifiable plant fossil (compression flora)
drifted plant stem
planar stratification
ripple cross-stratification
climbing ripple lamination
convolute / wavy lamination
planar cross-stratification
trough cross-stratification
load casts
slump
bioturbation
intraformational clasts
ca calcareous cement
sand dike
SH 85 section subjected to detailed petrographic & palynological analysis
xi sample interval
P isolated palynology sample
bright COAL (vitrain, clarain)
semi-bright COAL
intermediate COAL
clarain
semi-dull COAL (clarain, clarodurain)
dull COAL (durain)
shaly, impure COAL
coaly SHALE
MUDSTONE
(thin bed)
SILTSTONE
Fine grained SANDSTONE
Medium grained SANDSTONE
Coarse to conglomeratic SANDSTONE
erosional base
non-erosional or undifferentiated base
natural gamma ray log
gamma-gamma (density) log
decreasing γ counts
decreasing density
metres
feet
10
30
20
5
10
0
500
feet
50
100
200 metres
Sedimentology & Macroscopic Petrography of the No. 3 Seam, Springhill Coalfield
Vertical exaggeration 25:1
1989
J.H. Calder

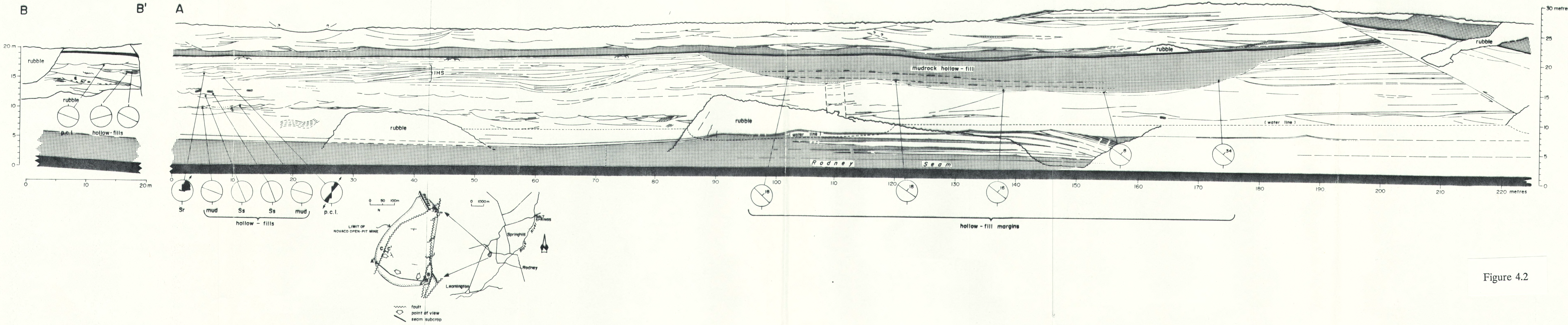
Figure 2.7

STRATIGRAPHIC CORRELATION OF BOREHOLES
SPRINGHILL COALFIELD
CUMBERLAND BASIN
J. H. Calder
1991
SOUTH
NORTH
RAGGED REEF FORMATION
HARRISON BROOK MEMBER
(Lithofacies assemblage VI)
MacCARRONS CREEK MEMBER
(Lithofacies assemblage V)
COAL MINE BROOK MEMBER
(Lithofacies assemblage IV)
LEAMINGTON MEMBER
POLLY BROOK FORMATION
SPRING HILL MINES FORMATION
JOGGINS FORMATION
(Lithofacies assemblage III)
INTERCALATED JOGGINS AND SPRINGHILL MINES FORMATIONS
(assemblage III & IV)
(Lithofacies assemblage II)
No. 3 SEAM (No. 3 MINE, SYNDICATE MINE)
No. 1 SEAM (No. 1 MINE)
No. 2 SEAM (No. 2 MINE, ABERDEEN SLOPE)
No. 7 SEAM (No. 4 MINE)
No. 7 SEAM (No. 7 MINE)
No. 6 SEAM (No. 4 MINE, No 6 Mine)
ROONEY SEAM
McCARTHY SEAM
No. 3 LOWER
DATUM
No. 3 SEAM
No. 2 SEAM
No. 2 UNDER
No. 5 SEAM
No. 7 SEAM
No. 6 SEAM
GESNER SEAM
Section highly faulted
minor red beds
red beds
(Correlations of Numbers 1 and 2 seams north of SH 12 are tentative)
LEGEND
SPRINGHILL
0
1 km
0
50
100m

Figure 3.1

LEGEND
climbing ripple lamination
planar / rythmic lamination
trough cross-lamination
slumped strata
convolute laminae
intraformational clasts
roots
identifiable plant compression
plant stem
pyrite nodules
coal
thin mudrock
silt
sand
conglomerate
paleoflow indicator (north at top)
palynology sample
bivalve fauna
mud silt v. fine fine medium coarse v. coarse pebble cobble
sandstone
conglomerate
RED
(early) Westphalian B
(late) Westphalian A
Westphalian A-B boundary ?
Spore zone VI
Spore zone IV B
Spore zone IV A
Spore zone III
(sampled in laterally equivalent drillholes)
No. 3 seam
No. 2 seam
No. 7 seam
No. 6 seam
Gesner seam
POLLY BROOK FORMATION
RAGGED REEF FORMATION
first appearence of grey sideritic mudrock
section is less red toward the south - coal seams being more abundant (noteably Rodney and McCarthy seams).
Drillhole SH 5
Drillhole SH 99
SH 73
Drillhole SH 74
SPRINGHILL MINES FORMATION
JOGGINS FORMATION
CUMBERLAND GROUP
interfingering of Springhill Mines, Polly Brook and Joggins Formations
500 m
100 m
50 m
0 m
? Westphalian A-B undifferentiated
strata concealed
exposed on South Polly Brook
exposed on Polly Brook east to River Phillip (R.J. Ryan, pers. comm.)
POLLY BROOK FORMATION
CUMBERLAND GROUP
(contact not exposed; probable fault contact here and coincidentally with apparent angular uncomformity on Black River at Deep Brook)
(early) Westphalian A
Spore zone II
ca
strata concealed
Section exposed on Deep Brook, Highway 142
BOSS POINT FORMATION
CLAREMONT FORMATION
CUMBERLAND GROUP (previously Riversdale Group)

Figure 3.3



B
B'
A
A'
rubble
hollow-fills
p.c.l.
20 m
15
10
5
0
10
20 m
IHS
mudrock hollow-fill
(water line)
water line
Rodney Seam
Sr
mud
Ss
Ss
mud
p.c.l.
hollow-fills
hollow-fill margins
30 metres
25
20
15
10
5
0
30
40
50
60
70
80
90
100
110
120
130
140
150
160
170
180
190
200
210
220 metres
LIMIT OF NOVACO OPEN-PIT MINE
fault
point of view
seam subcrop
Springhill
Salt Springs
Rodney
Leamington

Figure 4.2