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Paleogeography and Sedimentation of Late
Pleistocene Sand Deposits in the Upper
Ottawa Valley

By

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ABSTRACT

A major sand deposit in the

region in the late Pleistocene

of the Ottawa Valley

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ABSTRACT

Five major sand deposits in the upper Ottawa Valley have generally been interpreted in the literature as having been built by late Pleistocene glacial Lake Algonquin drainage--the Fossmill Outlet--which discharged into the Champlain Sea. This interpretation is supported by circumstantial evidence, however, no attempt has been made to substantiate a deltaic origin for the sand deposits with field evidence.

Paleogeographic analysis intergrated with a field survey and laboratory analysis of sedimentary characteristics of 33 sediment samples collected from sand deposits provides information which is used to establish paleoslopes and designate depositional environments associated with specific sand deposits by (1) trend surface analysis of sand deposit surfaces and (2) bivariate graphic analysis of moment measure values calculated from grain-size distributions of the sediment samples.

Although deltaic sedimentation processes were found to be important in the development of four of the sand deposits, the fifth, located in the 'Petawawa sand plains region' near C. F. B. Petawawa, Ontario, is interpreted as a littoral deposit developed in the Champlain Sea at the marine limit. The sediment supply to the sand deposits is shown to have originated principally from wastage of the Laurentide ice sheet; from which melt-waters transported sediments to the four deltaic sand deposits. Glacial lake overflow through the Fossmill Outlet was possibly of relatively minor importance as a source of discharge and sediments.

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...the value of knowing the genesis and geometry of a sand body precisely is at least twofold: that segment of geologic history and paleogeography represented by the sand is rather clearly known and the exploration and exploitation of natural resources in the sand may then be conducted on a sound technological basis involving less economic risk.

J. W. Shelton 1973

Sand deposits along the western bank of the Ottawa River between Bonnechere River and Deep River, Ontario, cover approximately 260 Km^2 (100 M^2) (map 1). The terrain of this part of the Ottawa Valley is dominated by the presence of these sands.¹

Sands that lie below approximately 150 m (490 ft) present elevation² have been partly eroded by fluvial entrenchment. Those sands are, for the most part, associated with terraces and paleochannels of the Ottawa River (Hanley 1972, French and Hanley 1975). However, sand deposits that lie above approximately 150 m appear as level or slightly rolling plains which were relatively spared from fluvial erosion. Sands above approximately 150 m in the Petawawa area (Jorgens and Airport sand deposits), the Indian River Valley (Alice and Pembroke sand deposits), and the Bonnechere River Valley (Bonnechere sand deposit) (map 1) have been interpreted as remnants of proglacial deltas (Mackay 1949, Chapman and Putman 1951, 1966 and 1973, Burger 1967, Prest 1969, Harrison 1971).

The deltas are generally interpreted in the literature as having been built by Pleistocene glacial lake drainage--the Fossmill outlet system³--which discharged into the Champlain Sea⁴ at approximately 170 m

¹The full range of sediment sizes, from boulders to clay, are found in the sand deposits; however, sand (defined by Wentworth (1922) as grain sizes ranging between 2.00 mm and 0.625 mm) is the predominant sediment type. In this thesis the term 'sands' refers to a sediment inclusive of all sediment sizes, unless otherwise stated.

²See page 3 for footnotes 2, 3, and 4.

(550 ft) between 12,800 and 10,000 years ago. This drainage system carried waters eastward from the ice-marginal Lake Algonquin⁵ to the Champlain Sea.

Based on this set of interpretations the thesis task was set initially as the development of a computer simulation of the deltaic sediments. Preliminary map, air photo and field analysis revealed that, in fact, topography and sedimentary variability of the deltas was too great to allow further progress on this task at present. Indeed, when reviewed from the perspective of this postponed task, the previous studies were seen neither to substantiate nor discredit the above interpretations because the lines of evidence used in support are mostly qualitative and speculative. A deltaic origin for the sand deposits has been deduced by previous studies on the basis of a number of external factors: geographic and stratigraphic position of the sand deposits in relation to traditional static models of regional stratigraphy; environments of deposition; pro-glacial lake outlets, and marine limits. No attempt has been made to substantiate a deltaic origin with field evidence, or other forms of quantitative information; indeed, the present study is the first to present a comprehensive statement on the sequence of geomorphic events leading specifically to the deposition of the sand deposits named above.

This statement is in the form of a paleogeographic reconstruction of the Ottawa Valley for the time of sand deposition. The reconstruction depicts those elements considered essential by Potter and Pettijohn (1963): the distribution of land, ice, and sea; basins of sedimentation; sediment dispersal systems; the position of shorelines; the source lands that provided the sediments which filled the basins.

The paleogeographic reconstruction developed in the following chapters is a major working hypothesis that was induced from evidence available from previous studies, field observations, sediment analysis, and computer assisted analysis of topographic maps.

In Chapter Two are presented: the relationships between regional geologic history and the sources and dispersal of sediments in the map area are analysed. Trend surface analysis is used to test the validity of the earlier hypothesis (summerized by Chapman and Putman 1973) that the Jorgens

and Airport sand deposits are remnants of a contemporaneous deltaic depositional surface with sediments supplied by the Petawawa and Barron Rivers.

Detailed information is not available from earlier studies on (1) the intrinsic sedimentary characteristics of the sand deposits or (2) the areal variation of sand deposit composition and structure. But this information is required to aid recognition of (1) the spatial and chronologic arrangement of sand deposits and (2) principal sediment dispersal and depositional patterns characterized by areal variations in sediment grain-size, sorting, and primary sedimentary directional structures. A field survey was designed to obtain the above information. Computer and graphic analysis of grain-size information of sediment samples collected from the sand deposits are used in Chapter Three as aids in interpreting depositional environments and sedimentary processes associated with the samples. Sedimentary models are proposed for sand deposits in the upper Ottawa Valley.

A proposed regional deglaciation is introduced in Chapter Four as a possible working hypothesis and is based primarily on evidence in the literature. A summary statement is given in Chapter Four on the spatial and chronologic arrangement of sedimentary events leading to the deposition of the sand deposits.

²Elevations given are present day elevations, unless otherwise stated.

³Fossmill outlet system is comprised of previously interlinked proglacial channels which are interpreted in the literature to have carried glacial Lake Algonquin overflow eastward across the Algonquin Highlands to the Ottawa Valley during late Pleistocene time.

⁴Champlain Sea is a marine water body which occupied the St. Lawrence-Ottawa Lowlands subsequent to the last deglaciation of the Lowlands.

⁵Lake Algonquin is a fresh-water ice-marginal lake which occupied the Huron and Michigan Lakes basins during late Pleistocene time.

CHAPTER TWO

ANALYSIS OF SAND DEPOSIT MORPHOLOGY

2.1 Exogenetic factors

The survival of bedrock controlled drainage systems, (Map 2) even though ice sheets flowed across the region, indicates that only details of the pre-glacial terrain were modified by the ice.

Map 2

Structural Provinces of Eastern Canada
and Adjacent United States

1. Laurentian Highlands	6. Adirondack Mountains
2. Ottawa-Bonnechere Graben	7. Adirondack Axis
3. Madawaska Block Mountains	8. Huron-Michigan Basin
4. St. Lawrence Basin	9. Algonquin Arch
5. Frontenac Arch	(after Kay 1942, and
Prest 1969)

pre-glacial terrain were modified by the ice (Ambrose 1964, Flint 1971). However, even superficial abrasion and localized gouging and quarring were sufficient to provide a considerable amount of rock debris to the basal ice zone. This is evident from the till sheets deposited over the map area (Gadd 1963, Harrison 1971). Attrition of rock debris to smaller and smaller pieces during transport in basal ice is well documented (Boulton 1971). Studies by Blatt (1967) indicate that the average grain size of quartz particles weathered from gneisses is 0.48 phi (1.4 mm)

with a standard deviation of 1.56 phi.⁸ According to Dreimanis and Vagners (1971), however, the terminal grain size of quartz from gneiss and granite rock types is 2.5 phi (0.18 mm). Quartz sands which cover a large part of the study area range between the above grain sizes (table 1 p.91)

Typically, till in the graben is sandy and non-calcareous (Gadd 1963, Burger 1967), but the full range is present in grain sizes from large boulders to clay. In the Mattawa-Nipissing area clay-sized particles account for less than 3 per cent and sand or greater-sized particles account for more than 75 per cent of till sediment (Harrison 1971). It is apparent from the high sand content of tills that the southerly flowing ice sheets carried, from the Precambrian Laurentian highlands (map 2), debris loaded with crystalline rock fragments; much of which was sand-size. Many of the large clasts in the till in the graben were probably locally entrained and moved perhaps only a few kilometers (Burger 1967). Ambrose (1964, p. 817) states "...the principal modification [of the post-faulting topography] for which the glaciers were responsible was to bury, in some places, the old topography in glacial debris". The ice sheets as geomorphic agents, in effect, served only to redistribute detritus from highlands to lowlands.

The last deglaciation of the map area occurred between approximately 12,500 years and 10,000 years ago (Gadd 1971, Harrison 1971). Large volumes of glacial debris and sediment-loaded melt-waters were dispersed from the receding ice sheet margin during that period. The pattern by which water-borne sediments were distributed in the Ottawa-Bonnechere graben reflects the pre-Pleistocene drainage system which is closely correlated to regional topography.

⁸Phi units are $-\log_2$ of the grain diameter in mm (Krumbein and Pettijohn 1938).

2.2 METHODOLOGY

2.2a Introduction

The area shown on map 4 is commonly referred to as the 'Petawawa sand plains region' (Chapman and Putman 1973). Map 5 is a computer calculated and drawn contour map of the Petawawa sand plains region (as shown in Map 4). This computer map accentuates the topography of the area. In this area most surficial sediments below approximately 160 m are sand and gravel and relatively small areas of till, clay and bedrock (Gadd 1963, Soil Survey Staff 1964, Chapman and Putman 1973). Sediments that lie below approximately 150 m are mostly associated with terraces and paleochannels of the Ottawa River (Hanley 1972, French and Hanley 1972)(map 6). A distinct fluvial terrace, with an upper surface at approximately 150 m runs across the map area subparallel to the Ottawa River (maps 4, 5, and 6).

The Airport and Jorgens sand deposits occupy most of the map area west of the Ottawa River between 159 m and 150 m. The surfaces of these sand deposits are level or slightly rolling plains which appear on air-photographs to be relatively lacking in fluvial channels, dunes, or other features that would indicate reworking of sediments subsequent to their formation. Post-Champlain Sea drainage evolution in the upper Ottawa River Valley is characterised by continual entrenchment of the Ottawa River into the unconsolidated valley fill (Hanley 1972). Most of the original fill has been removed and transported by the Ottawa River down valley (Chapman and Putman 1973). The 150 m (490 ft) terrace marks approximately the earliest level of the Ottawa River. This terrace is in part constructed of bedrock (Gadd 1963); which indicates that development of the terrace

Map 4: Topographic map of the 'Petawawa sand plains region' analysed by trend surface analysis. The boundary of this map area coincides exactly with those of Maps 5, 6, 8a, and 8b. This boundary encompasses the Jorgens sand deposit and the Airport sand deposit.

Scale 1:120,000 (original scale 1:50,000)

Map 4: Topographic map of the 'Petawawa sand plains region' analysed by trend surface analysis. The boundary of this map area coincides exactly with those of Maps 5, 6, 8a, and 8b. This boundary encompasses the Jorgens sand deposit and the Airport sand deposit.

Scale 1:120,000 (original scale 1:50,000)

scale 1:120,000

SYMBOLS

(original scale 1:50,000)

MINIMUM MAXIMUM (feet)	ABOVE										
	BELOW	420.00	440.00	460.00	480.00	500.00	520.00	540.00	560.00	580.00	600.00
LLLLLLLL		111111									
LLLLLLLL		111111									
LLLLLLLL		111111									
LLLLLLLL		111111									
LLLLLLLL		111111									

Map 5: Computer computed contour map. This map displays elevation data by interpolating a continuous surface in the regions where there are no data points, basing these interpolated values upon the distances to and the values of the neighboring data points. 997 data points were digitized from the source map (map 4). Points were taken along selected contour lines. Computed space coordinates and elevations of individual data points were recorded on computer cards for analysis by SYMAP and PREVU mapping programs (Carleton University Computing Services) (see Appendix for details).

Map 6: Computer computed 3-D map of the area shown in Maps 4 and 5. The same 997 data points used by SYMAP* to compute Map 5 were used by PREVU* to compute Map 6.

*see footnote 18, and Appendix , p.99).

Jorgens and Airport sand deposits are shown to be important geomorphic elements of the west side of the Ottawa Valley north of Pembroke.

Map 6: Computer computed 3-D map of the area shown in Maps 4 and 5. The same 997 data points used by SYMAP* to compute Map 5 were used by PREVU* to compute Map 6.

*see footnote 18, and Appendix , p.99).

Jorgens and Airport sand deposits are shown to be important geomorphic elements of the west side of the Ottawa Valley north of Pembroke.

may have been bedrock controlled. The sand deposits that lie above the terrace may have, then, originally extended eastward in the area now occupied by fluvial paleochannels and the Ottawa River.

It would seem from the topographic nature of the area and the spatial distribution of the sands that during the period of deposition, at the mouth of the Petawawa-Barron Rivers drainage channel, sediment particles were: (1) deposited over an area which extends approximately equal distances north and south of the channel mouth; (2) deposited not as a thick wedge (thickening basinward) as though the sediment partly filled a deep dish-shaped basin, but, rather as a relatively thin sheet which conforms closely to the pre-existing underlying topography.

The length and width of the deposit appears to have developed many times greater than its thickness. Evidence that this must have happened exists as two bedrock outcrops (approximately 155 m) (both nearly flush with the sand surface), one about mid-distance between the 150 m (490 ft) terrace and Radtke Hill (map 1), another an outcrop situated immediately at the brink of the 150 m terrace (Gadd 1963). These outcrops indicate that (a) the 13 m (40 ft) thick sand exposures located by Gadd (1963) along the face of the 150 m terrace between Petawawa and Chalk Bay may be only indications of the thickness of the sand deposit within the vicinity of the exposures, rather than the deposit being generally 13 m thick; (b) the depth of the Champlain Sea over the map area at the marine limit may not have exceeded 6 m (20 ft); and (3) sediments tended to move northeastward down the paleoslope.¹⁷

¹⁷"Paleoslope is at right angles to depositional strike which has usually been considered to lie subparallel to the strand line. The term paleoslope implies a subaqueous origin....the existence of preferred direction of cross-bedding in a deposit is independent of whether or not cross-direction reflects paleoslope" (Potter and Pettijohn, 1963, p. 82).

According to the above circumstances the general interpretation of the Airport and Jorgens sand deposits as being remnants of a single, contemporaneous deltaic depositional surface seems reasonable. Therefore, it was hypothesised:

- (1) the Airport and Jorgens sand deposits are remnants of a single, contemporaneous deltaic surface;
- (2) dispersal pattern of sediments was principally controlled by the paleoslope, which in turn, closely reflects the regional topography (Potter and Pettijohn 1963);
- (3) at any point on the hypothesised deltaic surface the surface slopes gently downward away from the Petawawa-Barron Rivers channel--the source of the sediments.

The validity of these hypotheses would be strongly supported provided that it could be demonstrated that the surface of the Airport and Jorgens sand deposits spatially coincide with a hypothetical surface in space which represents the paleoslope. The technique used to develop a 3-dimensional version of the paleoslope and to statistically measure the 'goodness of fit' of the sand deposits' topographic surfaces to the hypothetical surface was trend surface analysis.

2.2b Application of trend surface analysis

Trend surface procedures for analysis of geomorphologic problems are still in an early stage of development (King 1975). The mechanics of the method are given by Chorley and Haggett (1965), Harbaugh and Merriam (1968), Lustig (1969), Doornkamp (1972). However, trend surface procedures have not been applied previously to the type of problem analysed here but interpretation of the geomorphic relevance of the analysis results has been well established in the literature noted above.

A digital computer mapping program SYMAP¹⁸, was used to fit 3-dimensional polynomial surfaces to irregularly spaced point elevation data taken from the NTS map coverage of the study area. Seventy elevation points consisted of bench mark elevations and precise survey control

¹⁸ SYMAP is a packaged computer program, Carleton University, see Appendix for details pages 97-101.

points. Only 16 elevation points, however, that lie above the highest fluvial terrace (150 m (490 ft), Hanley 1972) were included in the analysis (map 7). The reason was as follows.

Entrenchment of drainage courses since deglaciation approximately 12,000 years ago has undoubtedly lowered the land surface over a large part of the study area. Only remnants of the hypothesised original depositional surface, such as the Airport and Jorgens sand deposits appear to have survived. As such, 54 of the 70 available elevation points are located on surfaces which are lower in elevation and probably post-date the development of the hypothetical deltaic depositional surface. The geomorphic significance of a trend surface fitted to all seventy elevation points, or at least to a data set which included many elevation points that are temporally and stratigraphically unrelated to the possible remnant surfaces, would be difficult to interpret: the development of the trend in the topography it would represent would be time transgressive. Doornkamp (1972) used similar criteria in collecting elevation point data for an analysis of planation surfaces in Uganda.

There is no advantage in using high degree trend surfaces where improvement in R^2 over lower degree surfaces is small because as the degree of the model equation increases so does the amount of local deviation included into the calculation of the model's coefficients. As such, the model becomes more accurate in predicting surface elevation,¹⁹ but its ability to separate local deviations from the regional trend decreases (Chorley and Haggett 1965).

The analysis was carried out to test hypotheses that were initially conceived through a geomorphic analysis of map, airphotograph, and field information and a review of the literature. Doornkamp (1972) states,

¹⁹ See Appendix for trend surface models expressed as equations page 97.

MAP 7

MAP 7

...this type of use of trend surface analysis can be justified but only on the grounds that:

- (1) the data conform as closely as possible to the conditions required for the valid application of the regression technique;
- (2) the data are based on accurately measured height data;
- (3) the hypotheses to be tested are based on geomorphological evidence in addition to the surface remnants themselves.

To meet the above conditions the following criteria were established for assessing the applicability of computed trend surfaces:

- (1) 90 per cent of the observed variance must be explained (R^2) (because "...the very purpose of the analysis is that all the regional trend should be removed from the observations and contained within the computed surface. However, when the trend does not account for over 90 per cent of the observed variance spatial autocorrelation remains within the deviations" (Doornkamp 1972);
- (2) analysis of variance test (F-statistic)²⁰ for the significance of R^2 must be $p \leq 0.001$;
- (3) residuals must be 1 per cent of the observed elevations (approximately ± 1.5 m (5 ft)) if there were to be grounds for using computed surfaces for predicting elevations of the original surface in areas where it may have been removed by erosion (i.e., upper and lower confidence surfaces about the trend surface (not calculated by SYMAP) would have to be very close to the true regression surface (Krumbein 1963, Harbaugh and Merriam 1965).

2.3 ANALYSIS RESULTS

2.3a Total sample analysis

1st, 2nd, and 3rd degree trend surfaces calculated on all 16 data points did not meet the above validating criteria (table 2^{*}). Coefficients of determination (R^2) for the 1st and 2nd degree surfaces were less than 0.90. The 3rd degree surface was statistically valid; however, 22 per cent of the residual values exceeded 1 per cent of the expected elevation

²⁰F was calculated by REGRESS: a Carleton University computer program package.

values. The 3rd degree trend surface also did not topographically conform with the actual surfaces of the sand deposits. Trend surfaces of the 4th order and higher improve R^2 and residual values, but, the form of the trend surfaces becomes increasingly unrealistic compared to the actual topography. The sand deposit surfaces seem not to coincide with a hypothetical surface in space which represents the paleoslope. Consequently, the hypotheses which state the Airport and Jorgens sand deposits were at one time part of a single deltaic depositional system (surface) was not supported by total sample analysis.

2.3b Split sample analysis

On the basis that no spatial association between elevation points on either side of Petawawa River was found by total sample analysis, the large sand deposits that lie north and south of Petawawa River are hypothesised to be two discrete, partial paleodepositional surfaces that are related to individual paleoslopes.

Trend surface analyses were undertaken on two separate groups of elevation points. The Jorgens sand deposit area provided 10 elevation points (from the original 16 point sample) (data points lying north of the line A - A' on map 7). The remaining 6 elevation points are south of that line. The position of the dividing line A - A' coincides approximately with the northern edge of the Airport sand deposit.

The results of 1st, 2nd, and 3rd degree trend surfaces on data groups north and south of the line A - A' are given in table 2. All calculated trend surfaces, with the exception of the 1st degree surface on the 'south' data group (12.5 per cent of the residuals exceeded expected values by more than 1 per cent), met the above validating criteria. But, "...statistical significance must be appraised and the decision of the [trend] surface's relevance made on the basis of the field worker's knowledge" (Andrews 1970). An appraisal of statistical values and the geomorphic character of the trend surfaces led to the rejection for both data groups of 1st and 3rd degree trend surfaces, the reasons being;

- (1) not only did 2nd degree trend surface analyses of both data groups provide lower residual values and higher R^2 values than 1st degree analyses, but the 2nd degree surfaces better simulated the topography of the actual surfaces (map 4, map 5, fig. 1, fig. 2);
- (2) for both data groups 3rd degree trend surfaces were topographically unrealistic in spite of being statistically legitimate (table 2).

The 2nd degree trend surfaces for data groups north (map 8a) and south (map 8b) of line A - A' (map 7) are obviously topographically dissimilar. "Trough" and "minimax" patterns were calculated respectively (after Chorley and Haggett 1965, fig. 15). The strength of correlations of the calculated split sample 2nd degree trend surface models and the remarkable morphologic likeness between the actual upper Ottawa Valley surface (including the sand deposit surfaces) and the trend surfaces indicate that the two 2nd degree trend surfaces do indeed portray real topographic trends (paleoslopes).

Consequently the following possibilities were directly suggested by the results of trend surface analyses:

- (1) the Airport and Jorgens sand deposits may not have been at one time part of a single depositional surface;
- (2) the sand deposits north and south of the Petawawa River may not have developed contemporaneously;
- (3) more than one sediment dispersal system may have been involved in the transport and sedimentation of sands in the map area;
- (4) calculated 2nd degree trend surfaces are valid reconstructions of discrete paleoslopes in the upper Ottawa Valley.

Trend surface analysis did not provide answers, rather it suggested some of the questions that should be asked and the direction in which the investigation should follow to obtain the information required to interpret the sedimentation history of the sands in the upper Ottawa Valley.

Having established by trend surface analysis that the Airport and Jorgens sand deposits are associated with different paleoslopes, and there-

Horizontal scale 1:50,000

Vertical scale 1:1,200

Figure 1 : Cross-profiles of the real topographic surface and computed 2nd degree trend surface of the upper Ottawa Valley near Petawawa - Jorgens sand deposit. The 2nd degree trend surface best simulates the local topography.

Horizontal scale 1:50,000

Vertical scale 1:1,200

Figure 2 : Cross-profiles of the real topographic surface and computed 2nd degree trend surface of the upper Ottawa Valley near Pembroke - Airport sand deposit. The 2nd degree trend surface best simulates the local topography.

Map 8a (top): Second degree trend surface map computed and drawn by SYMAP (elective 38) for the area north of line A - A' on map 7. The map was computed from 10 elevation data points. The 'trough' shaped trend surface is a good reconstruction of the topography of this part of the upper Ottawa Valley, i.e., paleoslope, as well as the local topography of the Jorgens sand deposit.

Map 8b (bottom): Second degree trend surface map computed and drawn by SYMAP (elective 38) for the area south of the line A - A' on map 7. The map was computed from 6 elevation data points. The 'minimax' shaped trend surface is a good reconstruction of the topography of this part of the upper Ottawa Valley between Allumette Island and the Algonquin highlands south of Petawawa, i.e., paleoslope, as well as the local topography of the Airport sand deposit.

Map 8b (bottom): Second degree trend surface map computed and drawn by SYMAP (elective 38) for the area south of the line A - A' on map 7. The map was computed from 6 elevation data points. The 'minimax' shaped trend surface is a good reconstruction of the topography of this part of the upper Ottawa Valley between Allumette Island and the Algonquin highlands south of Petawawa, i.e., paleoslope, as well as the local topography of the Airport sand deposit.

		ABOVE									
MINIMUM	BELOW	480.00	490.00	500.00	510.00	520.00	530.00	540.00	550.00		
		480.00	490.00	500.00	510.00	520.00	530.00	540.00	550.00		
ft											
SYMBOLS	LLLLLLLL	000000	111111	222222	333333	444444	555555	666666	777777	888888	999999
	LLLLLLLL	000000	111111	222222	333333	444444	555555	666666	777777	888888	999999
	LLLLLLLL	000100	111211	222322	333433	444544	555655	666766	777877	888988	999099
	LLLLLLLL	000000	111111	222222	333333	444444	555555	666666	777777	888888	999999
	LLLLLLLL	000000	111111	222222	333333	444444	555555	666666	777777	888888	999999

Map 8a (top): Second degree trend surface map computed and drawn by SYMAP (elective 38) for the area north of line A - A' on map 7. The map was computed from 10 elevation data points. The 'trough' shaped trend surface is a good reconstruction of the topography of this part of the upper Ottawa Valley, i.e., paleoslope, as well as the local topography of the Jorgens sand deposit.

Map 8b (bottom): Second degree trend surface map computed and drawn by SYMAP (elective 38) for the area south of the line A - A' on map 7. The map was computed from 6 elevation data points. The 'minimax' shaped trend surface is a good reconstruction of the topography of this part of the upper Ottawa Valley between Allumette Island and the Algonquin highlands south of Petawawa, i.e., paleoslope, as well as the local topography of the Airport sand deposit.

		ABOVE							
MINIMUM		BELOW	480.00	490.00	500.00	510.00	520.00	530.00	540.00
MAXIMUM		480.00	490.00	500.00	510.00	520.00	530.00	540.00	550.00
ft									
SYMBOLS	LLLLLLLL	000000	111111	222222	333333	444444	555555	666666	777777
	LLLLLLLL	000000	111111	222222	333333	444444	555555	666666	777777
	LLLLLLLL	000000	111111	222222	333333	444444	555555	666666	777777
	LLLLLLLL	000000	111111	222222	333333	444444	555555	666666	777777

fore, they may be products of at least two independent sediment dispersal systems, it would seem the the Petawawa-Barron Rivers drainage may not have been the only supply of sediments to these deposits, as has always been assumed (see also section 1.2).

The need for a new interpretation is apparent; preferably based, at least, in part on field evidence. In view of the fact that very little is known about sediment characteristics of these deposits, a field survey was directed toward obtaining information on sediment characteristics which would aid recognition of the depositional environments associated with the sand deposits.

In the following chapter the depositional environments associated with sediment samples collected from the sand deposits are designated by means of bivariate graphic analysis of moment measures and analyses of grain-size frequency distribution curves. This information is used to develop hypotheses(models) on sedimentation processes associated with the sand deposits.

CHAPTER THREE

ANALYSIS OF SAND DEPOSIT SEDIMENTARY ENVIRONMENTS

3.1 OTTAWA VALLEY SEDIMENTATION IN GEOMORPHIC PERSPECTIVE

3.1a Previous concepts

It was shown above that sediments supplied from westerly source areas by proglacial drainage to the upper Ottawa Valley were deposited progressively further up valley, and not contemporaneously as it has generally been assumed. The sediment was deposited, therefore, time transgressively. It is remarkable that the time transgressive nature of sedimentation in the upper Ottawa Valley is rarely noted in earlier studies. Indeed, the model held by many of the process of sedimentation is that each deposit in the valley fill represents discrete environments which existed consecutively in time (Ager 1973). The unconsolidated sedimentary fill is generally described in the literature as a sequence of till, clay, gravel, and sand (fig. 3a).

The difficulty that arose from the traditional concepts was that the model of the process of sedimentation of the valley fill was formulated from information collected as it is seen at exposures and boreholes. Little concern was apparently given to the dynamic manner in which the sediments were deposited.

3.1b Adding the 3rd dimension to sedimentation

It is helpful to visualize sediment deposition in 3-dimensions in order to appreciate best relationships between process and form. The deltaic fore-set bed in figure 3b at B is older than the top-set bed at A (principle of superposition). The vertical sequences at A-B and C-D are virtually identical, but it is difficult to determine stratigraphically if deposition of the fore-set bed in figure 3b at B was contemporaneous with that at D. If lateral continuity of a deltaic depositional environment existed between B and D then sedimentation at both locations would have probably occurred contemporaneously: similarity between sedimentary layers of sand located at B and D is explained by having vertical sedimentation take place concurrently at B and D with the same stratigraphic results. But, the point in fact that a depositional environment can be displaced laterally (Garner 1974, p.273) allows for the possibility

Figure 3a: Idealized diagram of the unconsolidated sedimentary fill of the Upper Ottawa Valley: the 'classic' model (after Karrow, 1961, fig. 3). A is a deltaic topset bed and B is a deltaic foreset bed, sediment supply is off diagram to the right. Sand and gravel on till ridge are sorted from the till by lacustrine or marine wave wash activity.

Figure 3b: One-point perspective view of the 'classic' 2-D model of sedimentary fill of the Upper Ottawa Valley. A and C are deltaic topset beds. B and D are deltaic bottomset beds. Sedimentation at-a-site occurs vertically, first B then A. As the deltaic environment moves northward toward D and C foreset beds are deposited first and these dip slightly from D to B to be followed by topset bed deposition.

that vertical sedimentation could have occurred at B but not at D at any given time. Vertical sedimentation with similar stratigraphic results to those developed at B could occur later at D if the deltaic depositional environment was laterally displaced from B toward D. The component would result in cross-bedding of sedimentary layers between B and D that would slope in the direction of D, although, probably at a diminutive angle that would be unrecognizable in the field (Ager 1973). The deltaic sand layer at B and D would appear sedimentologically similar provided conditions in the depositional environment remained unchanged during displacement. However, the deposits would be diachronous.

Sedimentation at the margin of a receding ice sheet is one situation where lateral migration of a sedimentary environment takes place. The layers of sedimentary fill in the Ottawa Valley may have, in effect, transgressed over the till-mantled Precambrian terrain in the 'wake' of the receding Laurentide ice sheet.

The implications of time transgressive sedimentation to sand deposits in the upper Ottawa Valley are dealt with below in the interpretation of field evidence of fluvioglacial and glaciomarine sediments that were deposited during ice sheet recession north of the Indian River.

3.2 METHODOLOGY

3.2a Introduction

Knowledge of sedimentary form and process inter-relationships was given little attention in previous interpretations of the sedimentary history of the upper Ottawa Valley sand deposits. These studies tended to lean heavily upon the idea of sedimentation as a two-dimensional phenomena in landscape genesis (see section 2.1). The contention is established above, however, that sedimentation processes should not be only looked at with respect to "top and bottom", but also with reference to "fore and aft" (Potter and Pettijohn 1963). Obviously, addition of the third-dimension to this study's perspective of sedimentary form and process inter-relationships strengthens insight into the genetic importance of various sedimentary characteristics recognized in the sand deposits.

In this chapter, computer and graphic analyses of grain-size information of sediment samples²¹ are used as aids in interpreting depositional environments and sedimentary processes associated with the samples. Sedimentary form and process models are proposed for the Airport, Jorgens, and Pembroke sand deposits.

Collection of sediment samples from the Bonnechere and Alice sand deposits was not possible, due to time, financial, and logistics limitations on this study. Evidence applicable to the interpretation of the

²¹See Appendix for sediment sampling and grain-size measurement procedures page 102.

formation of the Bonnechere and Alice sand deposits is given in chapter one; a summary statement on the formation of those sand deposits is given in chapter four.

3.2b Criteria applied in the interpretation of depositional environments

Standard bivariate plots of various combinations of the first four moments of grain-size distributions of sediment samples were used as aids in interpreting depositional environments and processes associated with the sediment samples. These techniques allow assessment of the degree of spatial homogeneity of sediments within and between the large sand deposits in the upper Ottawa Valley.

A computer program developed by Nickling (1972) was used for statistical computations (see Appendix for listing). The formulae used to calculate the moment measures are defined as follows (after Friedman 1967):

(1) Mean grain-size (in phi units) is $\bar{X} = 1/100 \sum f m$

(2) Standard deviation (a measure of grain-size sorting) is

$$Sd = (\sum f (m - \bar{X})^2 / 100)^{-1/2}$$

(3) Skewness (describes the symmetry of the distribution curve: a normal distribution has a skewness of 0.00) is

$$Sk = (1/100) \sigma^{-3} \sum f (m - \bar{X})^3$$

(4) Kurtosis (measures the peakedness of the distribution curve by comparing the sorting at the middle of the curve to that at the tails of the curve) is $Kt = (1/100) \sigma^{-4} \sum f (m - \bar{X})^4$, where

m = midclass size in phi units

f = percentage of total sample weight in each size class

$\bar{X}$ = mean size of sample

Friedman (1967) and Costello and Walker (1972) among others have successfully used the above statistical parameters to differentiate between sediments from beach, fluvial, and eolian type depositional environments. "The ability of these parameters to do this lies in the fact that the frequency distribution is dependant upon sediment supply, the mode of transport and the competency of the transporting medium" (Nickling 1972 after Rees 1970).

Data points on bivariate plots of standard deviation (Sd - sorting) against (1) mean size ($\bar{X}$) and (2) skewness (Sk - symmetry) tend to form distinct groupings. Other combinations of parameters did not show a tendency to separate data into distinguishable groups. It should be noted that calculated kurtosis (Kt) values, which are strongly influenced by the nature of the 'tails' of the distributions, were probably influenced by the lack of a large enough size range to make between sample differences apparent on the graphs. Only graphs that show distinct point groupings are given (figs. 4 and 5).

The result that some combinations of parameters did not segregate data points into groupings is not uncommon in this type of analysis because the environmental 'sensitivity' of individual parameters and parameter combinations can vary depending on between-sample variation and the portion of the grain-size distribution used in the analysis (Nickling 1972).

In this study, because of the coarse grained nature of the samples and the generally negligible proportion by weight of fines (silt and clay commonly comprise less than 3 per cent of the samples), fines were not included in the computed sediment grain-size distributions. It was supposed that the fines were selectively sorted by high energy modes of transport and consequently were transported away from the sand deposits. This winnowing process is easily identified from grain-size distribution curves of those sediment samples analysed between -1.00ϕ and $+4.25 \phi$ when plotted on log-probability graph paper: the 'fine tail' is 'cut off' for sediments which had their fines content mostly removed before deposition²² (Church and Gilbert 1975)

The underlying assumption with the standard use of bivariate graphs of moment measure values is that under sedimentary conditions of one type of

²² See samples 22 to 25 in figure 10.

Figure 4: Grain-size versus standard deviation of sediment samples 1 to 33

SYMBOLS

□ Jorgens sand deposit

● Pit H - Airport sand deposit

▲ Pembroke sand deposit

○ Airport sand deposit (except Pit H)

■ Lemke Lake kame

▽ Ottawa River paleochannel

X Sand dune - Pembroke deposit

△ Ottawa River terrace 145 m

Figure 5 : Skewness versus standard deviation of sediment samples 1 to 33

SYMBOLS, see fig. 4. Friedman's (1961) discriminant curve (beach/river environments) separates beach-inshore sediments from glaciofluvial sediments, but fails to segregate fluvial sediments. There is some agreement, however, overall with environment group identification based, in part, on Friedman's (1962) sorting guidelines and his 1961 graphic (discriminant curve) guidelines.

depositional environment (e.g., fluvial), sediment supply will develop, in response to maintained sedimentary conditions, a preferred (equilibrium) grain-size frequency distribution (the characteristics of which are partly characterised by moment measure values). In the real world, however, fluctuations in sedimentary conditions within a type of depositional environment result in sediments showing a range of grain-size frequencies. But, the assumption behind bivariate graph analysis holds that this variability in frequency distributions should only result in a limited range in moment measure values, and therefore, samples collected from sediments associated with a common depositional environment should appear on bivariate graphs as a scattered, but still distinguishable, grouping of points.

However, after sample points are plotted, and groupings of points are isolated, the problem remains of identifying the sediment types and depositional environments associated with particular groups of points. The results of Friedman's (1961) bivariate graphic analysis of moment measures calculated from 267 samples including dune, beach, and river sands are commonly accepted as guidelines for identifying sediment types from bivariate graphs when the associated depositional environments are unknown. It must be taken into consideration, however, that Friedman selected samples for his comparisons which had size and other characteristics suited to well-sorted, medium to fine grain beach and dune samples: river sand samples with more than 5 per cent of grains in excess of $+1.00 \phi$ (0.5 mm) size, such as those samples available to the present study, were not plotted by Friedman because "...river sands of coarse sand or gravel can be either positively or negatively skewed and no predictable relationship could be determined" (Friedman 1961).

Furthermore, Friedman (1961) found that although beach sands were dominantly negatively skewed seven per cent of his beach samples were positively skewed (the coarser beach sands of Friedman showed a positive skewness). Chappel (1967) experimented with bivariate plots of moment measures of beach and dune sands as a means of identifying Pleistocene strand lines. He found that dune sands which were positively skewed and

beach sands which were negatively skewed could develop opposite skewness after initial deposition because of the effects of variable winds mixing sediments and the introduction of new sediments during the time of development of any subsequent environment. Chappel (1967) restated Friedman's (1961) conclusion that the environmental 'sensitivity' of moment measures depends on whether the sediment was in equilibrium with the depositional environment and its composition not modified since initial deposition.

Proglacial sedimentation in the upper Ottawa Valley has been shown to be time transgressive: the receding ice front was a principal factor in control of sediment distribution. Consequently, the frequency and magnitude of sedimentary events was possibly very variable, in the short term (Østrem 1975). It seems reasonable to suspect, therefore, that few sediments, during the period of deposition of the sand deposits, established 'equilibrium' frequency distributions prior to the onset of a subsequent depositional environment. Some sediment sample moment measures, therefore, when plotted on bivariate graphs, should be expected to be anomolous with Friedman's guidelines.

As such, the inclusion of any sample point in a designated depositional environment grouping is based not only on Friedman's guidelines and its location relative to other points but also on these following factors: grain-size cumulative per cent frequency distribution curve shape, and geographic and stratigraphic position of sediments.

3.3 INTERPRETATION OF DEPOSITIONAL ENVIRONMENTS ASSOCIATED WITH SAND DEPOSIT SEDIMENTATION

3.3a Beach-inshore environment

Facies²³ Al is a surficial fine, poorly stratified sand (fig. 6) which forms a laterally discontinuous stratigraphic unit of the Airport

²³A sedimentary facies is defined as a sum of all the primary characteristics of a stratigraphic unit. A unit is a result of deposition in a given environment, and thus possesses characteristics of that specific environment (Reineck and Singh 1973).

sand deposit. Moment measure values of samples 5 and 6²⁴ collected from facies A1 (map 7) are similar to those of dune sands analysed by Friedman (1961) and dune sands (sample 33) analysed as part of this study.

A surficial tabular cross-bedded, fine-to-medium sand facies (J1) covers the area on the Jorgens sand deposit lying approximately between the 150 m (490 ft) terrace scarp and the Trans-Canada highway (map 5). This facies (J1) is essentially an elongated sedimentary unit which has its length running subparallel to the inferred paleoslope (established in section 2.3).

Figure 6 : A surficial fine, poorly stratified sand facies (A1) exposed at sample site 5, Airport sand deposit.

²⁴ Sediment sample numbers coincide with sediment sample site numbers shown on map 7: one sample collected per site (exception: at sample site 18 samples 18 and 19 were collected from the same section).

At sample sites 17 and 18 on the Jorgens sand **deposit** (facies J1) (map 7) a cobble and pebbles were found at one meter depth, respectively. Their presence within the otherwise sandy facies (J1) indicates they may have been ice-rafted to those sites; thus the tabular cross-bedded structures exposed at sample site 18 (fig. 7) were possibly developed subaqueously.

Figure 7 : A surficial fine-to-medium, tabular cross-bedded sand facies (J1) exposed at sample site 18, Jorgens sand deposit.

Folk and Ward (1957), Friedman (1961 and 1967), Fuller (1961), among others found that the points of inflexion in grain-size distribution curves, when plotted on arithmetic-probability paper, separate the modes of sediment transport and persistently occur at approximately $+0.80 \phi$, $+2.00 \phi$, and $+4.25 \phi$. The common interpretation of a multimodal curve is that the $+0.80 \phi$ inflexion point separates sand sizes moved by bottom traction processes (bed load) and saltation processes, the $+2.00 \phi$ inflexion point indicates

a change in mode from primarily saltation to suspension and saltation processes, and the +4.25 ϕ inflexion point indicates a mode change to suspension of fines. The concave downward shape of grain-size frequency distribution curves for samples 17 and 18 (fig. 8) near +2.00 ϕ is interpreted in accordance with the above as representing a change in transport mode from saltation to suspension and saltation around the +2.00 ϕ inflexion point.

That the sediment from which samples 17 and 18 were collected (facies J1) is (1) approximately 99.5 per cent (by weight) finer than +2.00 ϕ ; (2) cross-bedded with ripple markings; (3) possibly subaqueously deposited; and (4) ~~it has a~~ mean size and sorting characteristics which coincide with modern and Pleistocene outer and inner barrier bars analysed by Hails (1967) ($\bar{X}$ ranged from +1.95 ϕ to +2.30 ϕ and S_d ranged from 0.20 ϕ to 0.45 ϕ) indicates that this sediment (facies J1) was deposited partly out of suspension in slowly flowing water (lower regime) (Allen 1968, Reineck and Singh 1973).

Sample 18 was collected from a sediment bed that is apparently typical of the section (facies J1)(fig. 7); sample 19 was collected, at the same site, from the bed (one of the occasional coarser beds) immediately lying below the bed from which sample 18 was collected. The distribution curve for sample 19 (fig. 8) shows inflexion points around +0.80 ϕ , +2.00 ϕ , and +4.25 ϕ . The inference can, therefore, be made that the sediment from which sample 19 was collected (facies J1) was transported to the sample site in stronger-than-average bottom density currents in an inshore environment.

Sediment samples collected from facies A1 (Airport sands) and J1 (Jorgens sands) form the beach-inshore environment group designated in figures 4 and 5. It should be noted that boundary lines drawn around environment groups in these figures are not discriminant boundaries, but only serve to isolate, on the graphs, samples of possible common origin. Facies A1 is interpreted as having developed in an eolian environment and facies J1 in an inshore coastal environment.

FIGURE 38: Cumulative percentage grain-size frequency distribution curves for sediment samples collected from A1 (Airport sand deposit) and J1 (Jorgens sand deposit) facies, and dune sediments (sample 33). These samples are interpreted as representing sands deposited in a beach-inshore environment.

FIGURE 9 . Cumulative percentage grain-size frequency distribution curves for sediment samples collected from A2 facies, Airport sand deposit. These samples are interpreted as representing sands deposited in a fluvial environment.

FIGURE 10: Cumulative percentage grain-size frequency distribution curves for sediment samples collected from A3 (Airport sand deposit) and J2 (Jorgens sand deposit) facies, and kame sediments (sample 10). These samples are interpreted as representing sands deposited in a fluvio-glacial environment.

3.3b Fluvial environment

Facies A2 (Airport sands), comprised of a fine-to-medium horizontally-bedded sand, is separated from an underlying, sandy, graded, horizontally-bedded gravel facies (A3) by a disconformity (fig. 11a). Sections in the Airport sand deposit have been recorded at eight gravel pits (map 7). Six sections are shown, and a correlation of facies between sections is given, in figure 12. The A2/A3 facies disconformity is clearly evident at the sections, however, facies A2 is absent at gravel pit H (fig. 12).

The frequency distribution curves for the coarsest two samples (11 and 14) (fig. 9) collected from facies A2 have distinctive +0.80 ϕ and +2.00 ϕ inflexions: fluvial sediment is commonly multimodal (Friedman 1967). These curves converge near +2.00 ϕ and both samples have similar distributions in their fine tails. The similarity in these curves below +2.00 ϕ is seen to indicate that the suspended sediment loads in the flows were nearly identical in terms of the percentage of the total load represented in size distribution. Divergence of these curves above +2.00 ϕ indicates that competence was greater toward the proximal end of the Airport sand deposit (from the inferred paleoslope previously established).

Samples 12 and 13 are also multimodal (fig. 9), but inflexion points are less well-established than for samples 11 and 14. The fraction coarser than +0.80 ϕ represents less than 5 per cent of the distribution, i.e., the traction mode may not have been well-established because of low velocities.

All samples included in the fluvial environment group in figures 9 and 10 : (1) were collected from either horizontally-bedded or tabular cross-bedded sands, which in each case contained some gravel (-1.00 ϕ) and that from 5 to 25 per cent of the samples (by weight) was greater than +3.00 ϕ (very fine sand or finer) agrees with reported grain-size characteristics of fluvial sands (Friedman 1967, Costello and Walker 1972); (2) meet Friedman's (1962) sorting guidelines for a fluvial sediment.

Figure 11a: Section at Pit A--Airport sand deposit. Disconformity marks boundary separating sandy gravel below (A3 facies) and sand above (A2 facies). Both facies are characterized by horizontal bedding with no strong cross-bedding. Bedding structure of A3 facies is shown in detail in figure 11b.

Figure 11b: Detailed view of imbricated clasts in A3 facies at Pit A. If clasts dip in the inferred upstream direction, then flow was from right to left. Grading, fining upwards, is shown.

3.3c Fluvioglacial environment

A surficial medium-to-coarse cross-bedded sand facies (J2) extends westward from approximately the Trans-Canada highway to a scarp cut into the base of Radtke Hill which is inferred to mark the local marine limit near 159 m (520 ft) (map 5). Sediment samples 22 to 25 (map 7) collected from facies J2 tend to have the fine tail of their frequency distributions curves 'cut off' at approximately +3.00 ϕ (fig. 10). Very fine sand, silt, and clay typically account for only approximately 3 per cent of the total sample weight. The absence of a well-established fine tail in glacial outwash was also observed by Claque (1975) and Fahnestock (1963). They measured between 1 per cent and 10 per cent fines content; the low percentage being attributed to selective washing out of the fines during transport of the outwash away from the ice margin.

Facies A3, A4, and A5²⁵ are shown in figure 12 and figures 13a, 13b, 14a, and 14b (see also facies A3 in figure 16a). A notable portion of the quartz content of samples collected from these facies (table 3) was in the coarse sand-size range and grains commonly had sharp edges. These sediments, therefore, may have been transported only a short distance during the time the quartz content was weathered from the parent rock and deposited at sample sites (see also section 1.2).

Although the +3.00 ϕ to +4.25 ϕ (fine tail) portion of the frequency distribution was not analysed for sediments collected from these facies (A3 and A4) it is obvious from their distribution curves shown in figure 10 that less than 10 per cent of these sediments are finer than +3.00 ϕ (fine sand). Thus, they show grain-size characteristics that are similar to those sediments collected from facies J2.

All samples collected from facies J2, A3, and A4 meet Friedman's (1962) sorting guidelines for moderately sorted glaciofluvial sediment. These samples comprise the fluvioglacial environment group in figures 4 and 5 (table 3).

²⁵No grain-size data are available for facies A5. The facies is characterised by apparent uniform texture and structure, and is assumed to be associated with fluvioglacial sedimentation because of its stratigraphic position.

TABLE 3
ASSOCIATED DESIGNATED DEPOSITIONAL ENVIRONMENTS, SAND FACIES AND
SEDIMENT SAMPLES

<u>DEPOSITIONAL ENVIRONMENT</u>	<u>SAND FACIES</u>	<u>SEDIMENT SAMPLE NUMBERS</u>
Beach-inshore	J1 A1 sand dune	17, 18 5, 6 33
Fluvial	A2 paleochannel	11, 12, 13, 14 16
Fluvioglacial	J2 A3 A4 kame -----	22, 23, 24, 25 1, 2, 3, 4, 7, 8, 9 15, 28, 29, 30, 31 10 27.
Ottawa River terrace	J2	26
Pleistocene Beach	J1	19
High-energy inshore current	P1	32
Fluvial (reworked by deflation)	J1	20, 21
Fluvioglacial (reworked by inshore marine processes)		

Figure 12: Sections at gravel pits in the Airport sand deposit with correlation and corresponding sediment sample moment measure values; and core auger sediment samples are correlated with facies at sections.

UNIT

- 1 Sand facies A1 - Beach-inshore environment: possibly eolian sands
- 2 Sand facies A2 - Fluvial environment: possibly late deltaic fluvial sands
- 3 Sand facies A3 - Fluvio-glacial environment: possibly deltaic top-set outwash sands
- 4 Sand facies A4 - Fluvio-glacial environment: possibly deltaic fore-set outwash sands
- 5 Sand facies A5 - Fluvio-glacial environment: possibly deltaic bottom-set outwash sands

S1, S5, S9, S12

are core samples

Section at Pit B is similar to that at Pit C; only Pit C is shown.

Section at Pit D is similar to that at Pit E; only Pit E is shown.

N.B. - no precise elevation or thickness data on facies are available; only relative facies thicknesses and elevations are shown.

Figure 12: Sections at gravel pits in the Airport sand deposit with correlation and corresponding sediment sample moment measure values; and core auger sediment samples are correlated with facies at sections.

UNIT

- 1** Sand facies A1 - Beach-inshore environment: possibly eolian sands
 - 2** Sand facies A2 - Fluvial environment: possibly late deltaic fluvial sands
 - 3** Sand facies A3 - Fluvioglacial environment: possibly deltaic top-set outwash sands
 - 4** Sand facies A4 - Fluvioglacial environment: possibly deltaic fore-set outwash sands
 - 5** Sand facies A5 - Fluvioglacial environment: possibly deltaic bottom-set outwash sands
- S1, S5, S9, S12** are core samples

Section at Pit B is similar to that at Pit C; only Pit C is shown.

Section at Pit D is similar to that at Pit E; only Pit E is shown.

N.B.-no precise elevation or thickness data on facies are available; only relative facies thicknesses and elevations are shown.

Figure 13a: Deltaic structures at the west side of Pit H shows top-set (A3 facies), fore-set (A4 facies) and bottom-set (A5 facies) bedding.

Figure 13b: Close-view of deltaic structures shown in figure 13a.

Figure 14a: Deltaic structures at the east side of Pit H; shows fore-set (A4 facies) and bottom-set (A5 facies) bedding.

Figure 14b: Close-view of deltaic structures shown in figure 19a.

3.4 INTERPRETATION OF SEDIMENTATION PROCESSES ASSOCIATED WITH THE FORMATION OF THE SAND DEPOSITS

3.4a Jorgens sand deposit

The approximate geographic limits of facies J1 and J2 were initially inferred in the field. The graphic plot of moment measure values against distance along the Brindle Road traverse (map 7) of the sand deposit in figure 20 gives support to the inferred boundary separating facies J1 and J2 by clearly designating that a change in trend in sand characteristics occurs approximately one kilometer west of the Trans-Canada highway. Sands lying west of the highway (facies J2) seem to have relatively uniform characteristics nearly up to the base of the 159 m (520 ft) scarp at Radtke Hill; but sands lying east of the highway (facies J1) show a tendency to become progressively finer, better sorted, and more negatively skewed further away from the highway.

One sample (26) is a moderately well-sorted and positively skewed beach sand, following Friedman's (1962) sorting guidelines. This sample (26) was collected near the base of the 159 m (520 ft) scarp; it is coarser, but better sorted than other samples (20 to 25) collected along Brindle Road, and positively skewed (see section 3.1b, p.48), whereas the other samples tend to be near normally skewed. The slope of the grain-size frequency distribution curve of sample 26 also supports a beach genealogy (fig. 15): the curve is unique in that at the $+0.80 \phi$ inflexion point slope increases toward the fine tail; 80 per cent of the sample lies above the $+0.80 \phi$ inflexion point; and 95 per cent of the sample lies above the $+2.00 \phi$ inflexion point. These grain-size frequency characteristics are interpreted as follows.

Martini (1975) noted that samples from earlier (higher) beaches of glacial Lake Algonquin in the Wasaga Beach area showed moment measures indicating environments lying between fluvial and nearshore deposits. However, late-stage beaches were more typical of modern shoreline deposits. He attributed the peculiar characteristics of the earlier beach deposits to: the combination of nearby sources of till and stratified drift sediment;

Jorgens sand deposit traverse from Trans-Canada highway to Radtke Hill

inability of the nearshore environment to rework the sediment effectively (but most of the initial silt and clay content was winnowed out); and the relatively short time during which water level remained at one elevation (i.e., equilibrium was not established before water level changed and the beach was abandoned). It is suggested that a similar environmental situation applies to the sands around the site of sample 26 leading to the development of the grain-size distribution characteristics noted above: sample 26, therefore, is an anomolous point on figures 4 and 5 .

In addition there is topographic evidence to support the above contention. The ridge into which the 159 m scarp was cut is "...probably a kame composed of poorly-sorted bouldery gravels...[these] gravel deposits along the margins of the Petawawa Plain [Jorgens sands] have lost their glacial topography as a result of post-glacial flooding of the area" (Gadd 1963). Sediment samples were not collected from the kame sediments bordering the sands, but sample 10 (fig. 10 and map 7), collected from ice-contact stratified drift in a kame near Lemke Lake may well represent the possible initial grain-size distribution of source materials near sample site 26. The grain-size distribution curve for sample 26 and sample 10 cross each other remarkably close to the $+0.80 \phi$ and $+2.00 \phi$ inflexion points. Accordingly, the initial sediment supply (sample 10) was 65 per cent gravel ($\bar{X} -1.00 \phi$) and 19 per cent very coarse to coarse sand (-1.00ϕ $\bar{X}$ $+1.00 \phi$). The reworked sediment (sample 26) contains 5.5 per cent gravel and 80 per cent very coarse to coarse sand. It would seem that competency of the bed traction forces at the beach site may have been capable of moving sediments at approximately -1.00ϕ , i.e., the size of the largest particles moved (Morisawa 1968).

Sample 26 contained 61 per cent more very coarse to coarse sand, but approximately only 3 per cent more medium sand ($+1.00 \phi$ $\bar{X}$ $+2.00 \phi$) and 1 per cent less fine sand or finer sediment ($\bar{X}$ $+2.00 \phi$) than the possible initial supply (sample 10). Selective movement of sand grains in those size fractions in the surf zone²⁶ is one possible explanation why very coarse to coarse sand was found to be much more abundant, and medium and

²⁶Coastal terminology after Ruhe (1975).

fine sand to be about as abundant as in the possible initial sediment supply. King (1966, after Fuller 1962) states that sands finer than $+0.80 \phi$ (the inflexion point), i.e., medium and finer sands, are often lacking from sandy sediment (as they are in sample 26) from shallow water (nearshore) from about 2 to 4 meters in depth. King suggests that this sand is moved either offshore or onshore from shallow water because "...it is the size that can be most easily picked up and transported by both wind and water". The possibility that surficial sands around sample site 26 may have been derived from both erosion of kame sediment and reworking of submerged outwash or till by shoaling waves in the surf zone

fits a sediment dispersal model developed by Sutton, Lewis and Woodrow (1974). In their model subaqueous nearshore sand sheets in Lake Ontario are formed from sands derived from shore bluffs (glacial and glaciofluvial sediments), submerged tills and local streams. They found that local streams and erosion of shore bluffs were not as important as sources of sand to the offshore sand sheets as is reworking of submerged till sheets by shoaling waves.

Other samples (21 to 25) collected from sand facies J2 are nearly identical to each other in terms of moment measure values and their grain-size distribution curves slopes (fig. 10). These samples meet Friedman's (1962) sorting guidelines for a fluvioglacial sediment. The virtual uniformity of these sediment samples and an inferred fluvioglacial origin are evidence that facies J2 (Jorgens sands) may be interpreted as an outwash plain.

The outwash sediment (facies J2) was probably submerged no more than 6 meters by the Champlain Sea at the inferred marine limit. Most of the outwash plain, therefore, was possibly within the surf zone, and probably reworked by shoaling waves and current activity. Pettijohn, Potter, and Siever (1972, p. 361, after Swift 1970) state that "...given a linear sediment source at the shoreline, a largely random movement of sand on the shelf, and a linear sink at the shelf break, as shown in figure [16], a net transport of sand across the shelf by some type of weakly anisotropic

diffusion process seems necessary....Donahue and others (1966, fig. 2) found some evidence to support the idea [that] sand, probably in part reworked from older relict sediments, becomes finer seaward".

Figure 16: Coastal transport of submarine sediment model, after Swift (1970).

Accordingly, it is suggested that most of the very coarse to coarse sand and some medium sand sorted from the outwash and kame sediments may have tended to move shoreward (bed traction and saltation modes, respectively) and some medium sand and most finer sediment, sorted from outwash and kame sediments may have tended to move seaward (saltation and suspension modes, respectively).

Also, sample 21 (fig. 15 and map 7) along the Brindle Road near the Trans-Canada highway is clearly similar in grain-size characteristics, toward the coarse tail, to samples 22 to 25; however, unlike the other samples it has a well-established fine tail; 43 per cent of the sample is finer than $+2.00 \phi$ (the inflexion point). Sample 21 seems probably too coarse to be eolian sediment (dune sands are typically finer than $+1.49 \phi$ (Friedman 1961)), and yet too poorly sorted to be beach sand (Friedman 1961). As the grains in this sample (21) are, for the most part, subangular, the sample is interpreted to be fluvioglacial sediment (from the same outwash sheet represented by samples 22 to 25) that initially lacked a fine tail, as do samples 22 to 25, but subsequent to deposition a quantity of fine to very fine sand was added.

Samples 17, 18, and 19 (facies J1, Jorgens sands) were interpreted as beach-inshore sands (may be outer or inner barrier bar sands, see section 3.2a).

While studying major depositional environments in Phillip Bay, Victoria, Link (1967) observed that shallow-water sand facies were confined to shallow water, high energy areas because strong surf and current activity winnowed away the fines, but with increasing depth the strength of reworking currents diminished and finer material increases. The bedrock topography underlying sand facies J2 (Jorgens sands) is such (see section 2.1a) that only east of the 150 m (490 ft) scarp is the bedrock surface at low enough elevations that the depth of the Champlain Sea could have increased much more than its 6 m or less depth over most of the area west of the 150 meter (490 ft) scarp. Evidently, facies J2 rests upon a platform-like bedrock surface which near the 150 m scarp begins to drop off rapidly into a major bedrock trench (presumably pre-glacial) at present partly occupied by the Ottawa River (the break in slope east of the 150 m scarp edge is possibly a feature of bedrock topography, which was adopted by a former channel of the Ottawa River). Ball (1967) proposed a model of sand dispersal (similar in concept to that described by Pettijohn and others (1972, see above)) on a nearshore, shallow-water marine platform (fig. 17).

Figure 17: Schematic diagram shows how sand (mostly finer than +1.00 ϕ) moved offshore by current activity and deposited where currents dissipated in deep water (after Ball 1967).

Figure 18 is a model, based on the concepts shown in figures 16 and 17 above, for the sedimentation processes that could have developed the sediment characteristics measured from sediment samples collected from facies J1 and J2 of the Jorgens sand deposit.

Figure 18: A suggested situation which may have led to sedimentation of the Jorgens sand deposit. Reworking of glacial outwash or till, that lies on a submerged bedrock shelf, by shoaling waves in the surf zone, sorts out medium sands and finer sediment. Medium sands tend to be moved shoreward by currents and are deposited in, or near, the beach zone; finer sediments tend to be moved seaward and fine sands and occasional medium sands are deposited near the shelf break, but fines are carried into deep water by currents of progressively declining competence. Primary sedimentation of J1 facies was possibly barrier bar formation. J2 facies is possibly reworked relict glacial or fluvio-glacial sediments.

3.4b Airport sand deposit

Composite in structure, this deposit is comprised of five sand facies which are arranged vertically. They are the end products of different sedimentation processes which operated within glaciofluvial, fluvial and beach-inshore depositional environments (table 3).

The sequence in which these facies developed is inferred from their stratigraphic order. The distinct disconformities between facies are obvious indications of where important changes occurred in sedimentation process. However, because other evidence of both spatial and temporal changes in sedimentation processes (e.g., lateral declines in sediment grain-size) are typically not as obvious as a disconformity, the results of the sediment grain-size analyses of sediments collected from the Airport sand deposit are used as aids in interpreting the nature and sequence of sedimentation processes which formed the deposit.

Sand facies A3. The results of the sediment grain-size analyses support the stratigraphic correlation between sediment sample sites shown in figure 12 and the hypothesis that the deposit is for the most part fluvio-glacial (table 3). Indeed, the morphologic, lateral and vertical textural and structural characteristics of the Airport sand deposit, described below, are similar to those characteristics of the Yana and Scott outwash fans (Malaspina and Scott Glaciers, Alaska, respectively) described by Boothray and Ashley (1975) and of sandur deposits in Baffin Island described by Church and Gilbert (1975).

On the proximal end of the deposit (paleoslope established in section 2.3b) at Pit A (fig. 12 and map 7) the section begins with facies A3: rudimentary horizontal sandy gravel beds, in which sediment size gradually, but consistently, grades, coarser to finer, upwards into rudimentary horizontal sandy pebble beds (fig. 11a). Rust (1975) proposes with Church (1972) and McDonald and Bonerjee (1971) that the most consistent structure in the gravel of proximal braided outwash is poorly defined, horizontal bedding; and individual clasts in this sediment are typically well imbricated (as they are in facies A3 at Pit A, fig. 11b). In facies A3 clasts dipping in

the inferred upstream direction correlate well with the previously hypothesised sediment supply: the Petawawa-Barron Rivers drainage. Upstream imbrication indicates clasts were densely distributed on the channel bed, so there was little chance for individual clasts to move far without being caught by protruding clasts (Potter and Pettijohn 1963). Facies A3 has the characteristics (at Pit A) of: (1) imbricated clasts; (2) lack of development of cross-bedding; (3) rudimentary horizontal gravel bedding; (4) and upward grading in grain-size. These indicate that aggradation took place in shallow, but high velocity flow (possibly upper regime) (Church 1972, Reineck and Singh 1973, Shaw 1975), which during the period of development of facies A3, at least near Pit A, appears to have successively declined in competence (after Church and Gilbert 1975).

The cumulative per cent grain-size distribution curve for sediments samples approximately in the middle of the facies A3 exposure (sample 15, Pit A) is nearly identical to the curve for kame sediments collected near Lemke Lake (sample 10) (fig. 10). Indeed, sediment samples collected from facies A3 at other sites (samples 1, 7, and 8) in the Airport sand deposit show similar curves. But, it is apparent from the curves that sediment texture in facies A3 is characterised by a decrease in the proportion of sediment coarser than -1.00ϕ (gravel) and a corresponding increase in sediment of sand-size in the inferred downstream direction. Competence of the current system to move all the sand fraction from the source to the distal end of the Airport deposit can be inferred from (1) the similarity in moment measure values between sample 15 (proximal, Pit A) and samples 1, 3, 4, and 9 (distal, Pit H, map 7); and (2) the grain-size characteristics of facies A3 at Pit A (sample 10) are similar to those found by Clague (1975) of glacial outwash samples collected from gravel pits in the Elk River Valley, British Columbia. In the Elk River outwash gravel to sand to fines ratios (by weight) ranged from 52:46:2 to 94:5:1. The distribution for fluvioglacial sediments at Pit A (proximal, sample 15, facies A3) was 66:31:3, but at the distal end of the Airport deposit, in Pit H, the ratios ranged from 21:78:1 (sample 1, facies A3) to 1:96:3 (sample 29, facies A4).

Sand facies A4 and A5. There is evidence in the locality of Pit H (fig. 12, map 7) (distal end) of sufficient depth of water for deltaic fore-set sedimentation (fig. 19). This evidence consists of tabular cross-bedded pebbly sand (facies A3) which rests disconformably upon large scale inclined planar and curved basal elements of the fore-set beds of pebbly sand (facies A4) which in turn rest disconformably upon horizontal beds of fine sand (facies A5) (figs. 12, 13a, 13b, 14a and 14b). These bed structures are interpreted as top-set, fore-set, and bottom-set deltaic beds, respectively (see also Flint 1971, p. 193, fig. 7-25).

Stratigraphic sequence of fluvioglacial facies. Formation of the stratigraphic sequence of fluvioglacial facies (A5 through A3) is explained as follows:

- (1) deltaic bottom-set, fore-set, and top-set sedimentation took place where sufficient depth of water was present at the mouths of distributaries (facies A5, A4, and A3, respectively;
- (2) aggradation of the beds of distributaries, possibly carrying shallow, upper-regime flow, on the surface of the deposit formed facies A3;
- (3) during seaward progradation of the delta front a distributary channel advanced over the fore-set beds (facies A4, Pit H), and as it did so, top-set beds of facies A3 continued to develop in the channel over the fore-set beds (fig. 20). Consequently, some of the sediment load, particularly the coarser fractions, was sorted out during aggradation of the advancing channel bed prior to its dispersal at the delta front. As a result, the sediments deposited as fore-sets (facies A4) are less coarse than sediments deposited as top-sets (facies A3) (Saunderson 1975).

Bed structures more than approximately one meter below the disconformity between facies A2 and A3 (Airport sands) were not exposed at gravel pit locations between Pits A and H (fig. 12) because the water table at those sites prevented excavation below approximately two meters of the surface. However, if the deltaic sedimentation model described above is applied, it is reasonable to suspect that fore-set and bottom-set beds (facies A4 and A5, respectively) may lie further below the exposed sections

Figure 19: 'Classic' concept of deltaic sedimentation. As channel flow enters standing water currents dissipate and competence declines; sediment particles are deposited primarily according to grain-size, i.e., settling velocity of grains.

A coarse sediments are deposited as top-set and fore-set beds.

B fines, that are carried seaward by surface flow beyond the prodelta slope, are deposited as bottom-set beds. These beds are overlaid by the advancing fore-set beds.
(after Jopling 1965)

at least toward the distal end of the deposit. That samples collected from facies A4 fore-set beds at Pit H have remarkably consistent moment measure values and similar grain-size distribution curves to, but slightly finer than, samples collected from facies A3 sediments. Thus it is argued that these sediment characteristics are diagnostically significant in terms of identifying deltaic fore-set sediments (facies A4) from core samples, where exposures are not available. Sediment samples 1 and 9 were collected

Figure 20: A3 top-set facies lies over A4 fore-set facies at Pit H. Top-set beds are formed by aggradation of distributary channel bed as the channel advances seaward over the fore-set beds.

by hand auger (see Appendix* for sampling procedures) at different locations (map 7), but both sites are within 1.5 Km of Pit H. On the basis of moment measure values and grain-size distribution curve characteristics, sample 1 and 9 (fig. 12) are correlated with facies A3 and A4, respectively. Considering that these samples (1 and 9) were collected in the vicinity of Pit H, it seems possible that sample 1 is top-set deposited sediment and sample 9 is fore-set deposited sediment (the top-set beds at

*
p. 102

sample site 9 may be either missing or they are overall less than one meter thick (the maximum sampling depth by auger).

Sand facies A2. The disconformity separating facies A2 from underlying fluvioglacial sediments, when interpolated between exposures, appears to form a nearly level plane from proximal to distal ends of the Airport deposit (approximately 159 m (520 ft)). The surface of facies A2, however, slopes concavely upwards from proximal to distal ends, with thickness tending first to decrease rapidly away the proximal end, but decreases only gradually toward the distal end (facies A2 was not identified in areas lying east of the Pembroke Airfield). Also, comparisons of $\bar{X}$ and grain-size distribution curves between sediment samples 14, 11, 13, and 12 (fig. 9) collected from facies A2 at sites lying successively further down the inferred paleoslope indicate that grain-size decreases downslope from medium sand to fine sand. This apparent downslope loss of competence as well as a downslope change in bed structures from horizontal bedding toward the proximal end to cross and ripple-laminated structures at the distal end of the facies indicates that flow regime declined from possibly upper to lower levels in the downslope direction (table 7 in Church and Gilbert 1975). The inference can be made, therefore, that either velocity decreased or depth increased, or both, in the downslope direction during formation of facies A2.

Accordingly, the formation of facies A2 is interpreted as follows. Aggradation of the deposit, as progradation of the delta front continued, possibly resulted in progressive reduction of channel slope on the deposit's surface (fig. 21). Evidence to that effect is seen in the vertical at-a-site change in sediment size in facies A3 at Pit A if theoretical hydraulic geometry relationships are applied: Flow velocity, competence, and sediment concentration (factors in control of grain-size distribution of the instantaneous sediment load (Wolman 1955, Leopold, Wolman and Miller 1965)), each being directly related to channel bed slope,²⁷ may decrease in response to decreasing channel bed slope, and as a consequence grain-size of sediments deposited on the channel bed may decrease through

²⁷ Approximates the slope of the energy gradient (Leopold et al 1965); see Appendix for relationships expressed as formulae (table 4).

time. Alternatively, successive decreases in grain-sizes of sediment comprising the sediment supply may result from: successive recessions of the presumed principal sediment source (i.e., the ice margin) away from the section site resulting in a progressive decline in flow velocity, competence, and sediment concentration at the section site producing a decrease in sizes of sediments deposited at the site. (However, lateral migrations of distributary channels over the section site would not result in a fining-upwards channel bed deposit, unless competence or sizes in the sediment supply were reduced between lateral passes of channels over the section site, such that previously deposited sediments were consistently coarser than newly deposited channel bed sediments).

It is suggested that the combined effects of progradation of the delta front and recession of the ice margin were: (1) a decreased angle of slope of the deposit's surface; (2) declining flow competence and

Figure 21: Hypothetical sequence of long profiles of a deltaic deposit. As the delta front progrades seaward surface slope successively decreases. A similar situation may be responsible, in part, for the successive decrease in sediment grain-size through A3 facies at Pit A.

grain-sizes in the sediment supply through time; (3) a fining-upwards sequence in sediments deposited at any given site on the deposit upslope from the delta front; (4) development of the A2/A3 facies disconformity when a critical threshold of competence was reached (The active gravel and sand

channel bed became a relatively inactive 'armoured' surface upon which aggradation of sands (facies A2) continued, apparently uninterrupted during the transition from predominantly gravel bed load to sand bed load).

By the time development of facies A2 was initiated the entire Airport sand deposit was probably emerging by isostatic processes: sea level was falling relative to the land. Under this circumstance aggradation in channels would have given way to entrenchment as base levels fell. That facies A2 appears to be restricted to areas lying west of the Pembroke airfield indicates aggradation within the fluvial system occurred only in the area approximately defined by the present-day Black Bay Creek catchment area. The depositional surface was then abandoned and entrenched by early, post-depositional drainages through the Black Bay Creek area.

Sand facies A1. A discontinuous bed of fine, very well-sorted sand lies upon fluvial sediments (facies A2) in some areas on the Airport deposit above approximately 160 m (525 ft). Moment measure values of samples 5 and 6 collected from this facies show on bivariate plots (figs. 4 and 5) that the sand was possibly deposited in a beach-inshore depositional environment. These same samples (5 and 6) have moment measure values similar to dune sands analysed by Friedman (1961) and as part of this study (sample 33). It seems, therefore, that sand facies A1 developed by eolian sedimentation, subsequent to the abandonment of the paleodepositional surface by the fluvial system, i.e., upon termination of facies A2 sedimentation.

Sedimentation model for the Airport sand deposit. The characteristics of facies A2, described above, and those characteristics of fluvioglacial facies A3 to A5 are consistent with a sedimentation model of an outwash delta proposed by Salisbury (1892: described by Church and Gilbert 1975). A sedimentation model of the Airport sand deposit, which is based in part on concepts developed by Salisbury is shown in figure 22. Thus, the Airport sand deposit is interpreted as an outwash delta resulting principally from proglacial outwash grading into the Champlain Sea.

AIRPORT SAND DEPOSIT

GRAVEL PIT SITES

Figure 22: Suggested morphometry and conditions of sedimentation of the Airport sand deposit.

Proglacial outwash sediments were supplied by melt-water channels that occupied the present Petawawa and Barron Rivers valleys. This sediment was first deposited in shallow water near the shoreline of the Champlain Sea at the marine limit. As the sediments advanced further seaward deeper water permitted the development of fore-set bedding as well as bottom-set and top-set bedding. The coarse grained sediments and lack of fines in the deposit indicates a high-energy depositional environment. A fining upwards of sediment grain-size through facies A3 and into facies A2 indicates progressive decline in sediment transport competence during the latter period of sedimentation of the Airport deposit. (after Salisbury 1892 in Church and Gilbert 1975)

(see figure 17 for a description of A1 to A5 facies shown in this figure).

N.B.- surface slopes and thicknesses of facies are shown in relative terms.

3.4c Pembroke sand deposit

The present surface of the Pembroke deposit is dissected by the deeply entrenched Indian River. On either side of the river the deposit surface extends as a nearly level plain. This surface lies approximately between 150 m (490 ft) and 152 m (500 ft).²⁸ Bedrock outcrops rise above the plain, and local sand dunes (apparently stabilized) are as much as 3 m high. Sample 34 was collected from a dune on the surface of the Pembroke deposit.

It was noted above (section 3.2a) that the Pembroke deposit appeared, at the only available exposure, to be relatively uniform in texture and structure; and therefore, only one sand facies (P1), was identified within the top 3 meters of the deposit.

Sediment analysis for sample 33 (fig. 8) collected from facies P1 shows no apparent resemblance to the sediment analysis results for samples collected from other sand deposits: moment measure values, and shapes of the grain-size distribution curves are very different from samples from other deposits. Also, sample 33 is too coarse to be considered dune sand and too well-sorted to be considered to be beach or fluvial sands if Friedman's (1961) guidelines are applied. The sand may be interpreted as marine sand below wave base if Friedman's (1962) sorting guidelines are applied. Burger (1967), however, shows the Pembroke deposit as being deltaic, but he gives no firm evidence to support that contention. Consequently, sample 33 is left as an anomalous point in figures 4 and 5.

The relatively gentle slope of the grain-size distribution curve (fig. 8) for sample 33, particularly through the medium sand fraction (+1.00 ϕ to +2.00 ϕ), indicates sediment transport predominantly in the saltation mode. Indeed, the sediment sample is 88 per cent (by weight) medium sand. Overall, sample 33 is 99 per cent sand coarser than +3.00 ϕ and 1 per cent sediment finer than +3.00 ϕ . It is evident from the downward inflexion of the curve near +2.00 ϕ (the inflexion point) that sediments finer than this (fine sand or finer sediment) were selectively removed from facies P1 sediment during, or subsequent to, deposition.

The grain-size distribution curve for sample 34 (fig. 8) collected from dune sand) has a well-developed 'tail' finer than +2.00 ϕ . The sediment sample (34) is, in fact, 80 per cent finer than +2.00 ϕ and medium sand comprises most of the sediment coarser than +2.00 ϕ .

On the scant evidence available it seems reasonable to agree with Burger (1967) that the Pembroke deposit had a deltaic origin. However, in addition, it is suggested that facies P1 sediments were possibly initially more typical of fluvial sediments in terms of grain-size frequency distribution, but subsequent to deposition and emergence, the deltaic depositional surface was subjected to deflation processes (Flint 1971, Reineck and Singh (1973). If competence of the wind was around +2.00 ϕ at the time, sediment finer than approximately +2.00 ϕ may have been entrained by the wind, and once air-borne, tended to be blown away from the deposit. But, sand grains around +2.00 ϕ and coarser in size were perhaps only moved short distances on the deposit. Thus, the proportion of total sediment comprised of medium sand and coarser sediment may have increased and the proportion of fines decreased to those shown by sample 33 (which is assumed to be representative of the upper 3 m of the Pembroke deposit).

SUMMARY AND CONCLUDING REMARKS

A paleogeographic reconstruction of major sand deposits in the Ottawa, Bonnechere and Indian River valleys shows that sedimentation of these sand deposits was complex; far more so than the overall deltaic genesis generally interpreted in the literature. The idea is developed that these sand deposits are a response to a relatively few major sediment dispersal patterns (Potter and Pettijohn 1963), i.e., sediment supply followed systematically arranged preglacial, fault-controlled routes which operated, in turn, in response to regional deglaciation patterns.

4.1 TOPOGRAPHIC AND PALEOHYDROLOGIC RELATIONSHIPS BETWEEN THE EASTERN SIDE OF THE HURON LAKE BASIN OCCUPIED BY LAKE ALGONQUIN AND THE OTTAWA-BONNECHERE GRABEN

...the premise here is that we build from things seen and analysed, however provisionally, to a comparison with data from elsewhere, from someone else, or inferred by necessity from a past that cannot be seen.

Carl O. Sauer 1956

4.2a A new look at old data on eastward draining outlets of Lake Algonquin

Glacial Lake Algonquin was bounded to the north by the receding Laurentide ice sheet. "The retreating [receding] ice exposed highland areas progressively farther to the north so that, to the waters of this lake, the strand is time transgressive" (Harrison 1971, p. 28). A radiocarbon date of $11,800 \pm 400$ years before 1950 (GSC-1363) for organic material collected from the bottom of a small lake by Harrison (1971) about 58 Km (20 M) north of South River, Ontario, indicates that deglaciation of the eastern side of the Huron Lake basin as far north as the Powassan-Fossmill area (map 1) may have begun by 12,200 years ago.

The northward decline of elevation of main Lake Algonquin raised beaches near Bernard Lake (380 m (1,245 ft)), South River, Ontario (372 m (1,220 ft)), and Trout Creek, Ontario (364 m (1,195 ft)) (Chapman 1954) (map 1 at D, E, and F, respectively) is attributed to the effects of differential uplift (approximately 0.26 m/Km (2.5 ft/M)), and not to lake

drainage through progressively lower easterly flowing lake outlets, by Chapman (1954) and Chapman and Putman (1973). Harrison (1971), however, attributes the decline in beach elevations northward of Bernard Lake to drainage of glacial Lake Algonquin overflow through a series of progressively lower lake outlets, opened by ice recession, to Mink Lake (map 1 at G), from where the waters entered the Petawawa River.

Harrison (1971, p. 27, fig. 16) constructed continuous water-plane profiles through the north to south extent of Lake Algonquin. He claims to have overcome some of the "...considerable difficulty encountered in correlating [discontinuous strand line data] over long distances, especially in the more rugged areas north of Lake Simcoe" (Harrison 1971, p.26) by using as data points sill elevations of specific glacial lake outlets in the North Bay-Mattawa area which he proposed as the northern extremities of the appropriate lake stages (fig.23).

The profile of strand line elevations of main Lake Algonquin (fig.23) when extrapolated northward of Bernard Lake gives a water-plane elevation of 402 m (1,320 ft) at South River and 445 m (1,460 ft) near Fossmill. These figures were used by Harrison to suggest that the 30 m (100 ft) difference at South River between the elevation of actual strand line evidence (South River beach 372 m) and the extrapolated strand elevation (402 m) was the result of lowering of Lake Algonquin by eastward draining overflow through a previously unreported lake outlet at South River (sill elevation 384 m (1,260 ft), Harrison 1971).

It would appear, if one accepts the beach site at South River (372 m) identified by Chapman (1954) as having been formed by Lake Algonquin, that initially there may have been lake overflow across Harrison's (1971) South River sill. However, another agency for lowering the lake level to the elevation of the South River beach site (372 m) obviously had to have operated. Also, the orientation of the topography is such that any lake waters that spilled over the South River sill had to have drained northward by way of Amable du Fond River (Waskigomog and Wilkes Lakes) for approximately 50 Km (15 M) along the west facing flank of the Amable du Fond-Nipissing Rivers drainage divide until a water gap was reached through which the lake waters could drain eastward toward the Ottawa-Bonnechere graben (map 1). Harrison's (1971) sequence requires that the ice margin be located north of Fossmill (the closest water gap) at that time, which specifically had that been the case then lake waters would surely have flooded the Huron Lake and Amable du Fond basins up to the ice margin, thereby putting Foss-

Figure 23: Strand line profiles drawn by Harrison (1971, fig. 16) of various levels of glacial Lake Algonquin. The construction of these profiles is based on the assumption that no isostatic recovery of the earth's crust took place while individual time transgressive strand lines developed. Profiles end in the north at lake outlets proposed by Harrison (1971) and Chapman (1954).

mill under approximately 128 m (420 ft) of water, as well as drowning the area occupied by the South River channel proposed by Harrison (1971).

Accordingly strand line evidence (fig.1) of Lake Algonquin should be found near 445 m (1,460 ft) at Fossmill. The highest evidence (indirect at that) of Lake Algonquin near Fossmill, however, is the Kilrush Lake sill (348 m (1,140 ft) (proposed by Chapman 1954, Harrison 1971) of the Fossmill outlet (which is now occupied west of Kilrush Lake by the Wistawasing River) (map 1, fig.24). Evidently the South River outlet channel was not responsible for lowering the level of Lake Algonquin by 30 m (100 m), i. e., between the levels of the Lake Bernard beach and the extrapolated lake level at South River. Instead, it appears more likely that the apparent 30 m lowering of the lake level was the result of a different sequence of events. In the following sections (4.2b, 4.2c, and 4.2d) evidence which has become available since Harrison's study is interpreted and an alternative sequence of events to earlier attempts is suggested for lowering of the lake levels and operation of lake overflow systems.

4.1b Recent evidence on lake levels and eastward overflow of Lake Algonquin

The following evidence helps to support Chapman's idea of isostatic rebound being a principal cause for the decline in strand line elevations north of Bernard Lake.

Dadswell (1974) pointed out that post-glacial redispersal of a group of fresh water crustaceans⁹ followed the receding ice northward toward the study area in glacial Lake Algonquin. "Their present distributions are restricted to lake basins in areas formerly occupied by the large interconnected glacial lakes and their spillways" (Dadswell 1974, p.49). The presence of crustaceans in some lakes up to, but not higher than, 381 m (1,250 ft) in the Amable du Fond and Petawawa River basins (map 1) indicates that glacial Lake Algonquin and these lakes were interconnected.

⁹The crustaceans sampled by Dadswell: Mysis relicta Loven, Pontoporeia affinis Lindstrom, Gammaracanthus loricatus Sabine, Limnocalanus macrurus Sars, and Saduria entomon (L.).

Long Profiles Fossmill Overflow Channel System

Figure 24: The channels shown occupy probably pre-glacial, structurally controlled valleys. Channels that lie west of the White Partridge Lake sill may have been initially submerged when glacial lake waters filled the Lake Algonquin and Lake Amable du Fond lake basins to the 386 m (1,266 ft) level. As the respective sills of these channels emerged glacial lake overflow may have flowed either westerly as well as easterly through the channels, i.e., for a time lake overflow may have drained in the opposite direction (westward through the Fossmill system west of Mink Lake) to the easterly direction that has always been assumed. Channels east of the White Partridge Lake sill carried glacial melt-waters eastward to glacial Lake St. Lawrence (Round Lake level) or the Champlain Sea (Percy Lake level). The discharge of these channels was possibly augmented by glacial lake overflow, that originated, first, west of the White Partridge Lake sill, then, after lake levels fell below that sill elevation, west of the Mink Lake sill. Easterly drainage of lake overflow possibly ended when the Mink Lake sill emerged.

Dadswell (1974) suggested that while the ice margin was situated between present day White Partridge Lake and Lake Traverse it dammed eastward draining proglacial melt-waters and Lake Algonquin overflow by plugging all Fossmill overflow channels (with sills lower than approximately 381 m (1,250 ft)) thereby creating glacial Lake Amable du Fond (map 1 and fig. 24). The Bonnechere River overflow channel sill (381 m) just southeast of White Partridge Lake (map 1) probably controlled Amable du Fond lake level (Dadswell 1974). The present day lakes, with surface elevations up to 381 m, in which Dadswell found crustaceans generally outline the area previously covered by glacial Lake Amable du Fond (Dadswell 1974, p.10).

The crustaceans sampled by Dadswell were light-avoiding and bottom dwelling types. In present day arctic regions these crustaceans are found only in lakes deeper than 5 m (16 ft) (Johnson 1962). It is assumed (with Dadswell 1974) that 5 m was the minimum lake depth required for crustacean survival in glacial lakes edging the Laurentide ice sheet. Thus, crustaceans found in lakes at 381 m (1,250 ft) in glacial Lake Amable du Fond basin can be accepted as evidence that the glacial lake may have stood as high as 386 m (1,266 ft) when the ice margin was near present day Koistkokwi Lake (map 1 at H).

The ice margin at that position would have permitted:

- (1) glacial lake waters to fill the Amable du Fond lake basin thereby creating a continuous water body that directly connected to the glacial Lake Algonquin water-plane south of Fossmill and extending as far south as the drowned South River channel (sill 384 m (1,260 ft));
- (2) crustaceans from Lake Algonquin to disperse eastward through Lake Amable du Fond to areas within the present day Petawawa River basin where populations exist in lakes with surface elevations up to 381 m (1,250 ft).

Lake Amable du Fond may have been, therefore, an extension of the Lake Algonquin water-plane as far eastward as the White Partridge Lake sill (386 m) (fig. 24). The fact that the line of apparent maximum tilt due

to differential uplift (Goldthwait 1908, Harrison 1971) and esker and striae patterns are oriented at right angles to a west-northwest to east-southeast line between White Partridge Lake and Koistkokwi Lake suggests that the component of regional differential uplift would have had a negligible effect on the water-plane (386 m) along the west to east axis of the lake.

A 30 m (100 ft) head of lake water at the South River (Harrison 1971) or White Partridge Lake sills following clearance of ice from respective outlet channels is unlikely if the fact is considered, in addition to the evidence given above, that no field evidence is given in the literature that indicates outflow of lake waters in catastrophic proportions occurred through the Fossmill outlet system. Indeed, it took over 50 years for researchers to find evidence of easterly lake outlets at the north end of main Lake Algonquin; and these outlets (Harrison's South River outlet and Dadswell's White Partridge Lake outlet) were identified on the basis of their geographic locations and elevations in relation to a high-stage water-plane interpreted either from extrapolated or indirect evidence and not on field evidence.

The South River beach (372 m (1,220 ft)) was not formed along a synchronous strand line (i.e., the main Lake Algonquin strand drawn and extrapolated by Harrison (1971, fig. 16) to derive an expected water level at South River at 402 m (1,320 ft)) because the South River outlet permitted the lake level to fall by at least 30 m (100 ft) as proposed by Harrison (1971); but rather the hypothesis is put forward that the decline in beach elevations north of Bernard Lake is a result of a more intricate system principally involving the related effects of ice recession, topography, and regional isostasy on lake configuration. An analysis of the proposed system is given in the following sections (4.2c and 4.2d).

Evidently the key factor in control of the main Lake Algonquin water level and easterly outflow was the position of the ice margin, especially where ice plugged the Petawawa River basin near White Partridge lake (Dadswell 1974) thereby closing the only major west to east overflow

route south of the Mattawa Valley. In the following section the position of the ice margin around the time of the closing of the main Lake Algonquin stage is reconstructed; and it is shown that during that period in glacial history two critical events occurred: one, the Champlain Sea was initiated in the Ottawa Valley, and the other, the first sedimentation of major sand deposits in the Ottawa-Bonnechere graben.

4.1c The position of the ice margin at the close of the main Lake Algonquin stage and initiation of the Champlain Sea

The ice margin at the initiation of the Champlain Sea in the Ottawa Valley (approximately 12,000 years ago, Gadd 1971, p. 92) was "rather definitely fixed" (Goldthwait 1933) by studies of fresh and marine stratified sediments between Ottawa and Montreal by Antevs (1925 and 1928). The ice margin ran approximately through Renfrew, Quyon, Carleton Place and south of Hawksbury and Montreal (Goldthwait 1933). The reconstruction is suggested also by the fact that "...the ice border ran nearly east and west, whereas, the striae, [for example drumlins and eskers mapped by Mackay (1949) and Chapman and Putman (1973)] follow a generally southerly course" (Goldthwait 1933).

Westward from Renfrew the ice margin probably continued up the Bonnechere River valley toward Eganville, building the moraine at the south side of present Mink Lake (map 1 at I) (Burger 1967, Chapman and Putman 1973). The writer proposes that waters of glacial Lake St. Lawrence¹⁰ may have flooded the present Golden Lake and Round Lake area by way of the Bonnechere River valley that was bordered by the ice margin just to the north (map 1). Evidence that the maximum elevation reached by Lake St. Lawrence in the Bonnechere River valley may have been approximately 183 m (600 ft) is given below. The proposed 183 m maximum lake level is in accordance with:

(1) Goldthwait (1933) stated "...the [210 m] 690 - foot sand deposit north of Kingsmere, Quebec, hitherto assumed to be marine because it is the highest and oldest watermark [reported by DeGeer 1892], lies on or

¹⁰ Lake St. Lawrence is a fresh water glacial lake which immediately preceded the Champlain Sea in occupation of the Ottawa-St. Lawrence Lowlands.

somewhat below the Lake St. Lawrence water-plane";

(2) published isostatic depression isobase maps for the Ottawa Valley by Prest, Grant and Rampton 1968, Map 1253 A and Dadswell 1974, Fig. 3A.

(3) Burger's (1967) observations that "...in the wide part of the Bonnechere Valley, in which Round Lake and Golden Lake are located, traces of an ancient shoreline occur at approximately 600 ft instead of between 500 ft and 520 ft in the Ottawa Valley. These traces of shoreline include wave-cut terraces, areas of bare bedrock and lag concentrations. Glaciofluvial deposits with crests just below 600 ft have a smooth washed-over appearance...; those just above 600 ft have sharp features";

(4) Rust and Romanelli (1975, p. 180, Table 2) inferred a marine limit at 198 m (650 ft) around 12,000 years ago in the Ottawa-Gatineau area on the basis of radiocarbon dates given marine shells found at Cantley, Quebec (12,000 $\pm$ 160 years, GSC-1646, 194 m (635 ft)). Evidently just prior to 12,000 years ago thinning of the ice sheet where it pressed against the north flank of the Adirondack Mountains near Quebec City permitted Lake St. Lawrence (210 m (690 ft) near Ottawa) to drain to sea level (198 m (650 ft)) (Gadd 1971, Flint 1971). A 12 m (40 ft) drop in water level may have also occurred at the same time in the Renfrew-Eganville area when Lake St. Lawrence at 183 m (600 ft) drained to sea level at 171 m (560 ft) with the marine limit near Renfrew identified by Goldthwait (1933) and Antevs (1925). A very rapid change from fresh-water to marine conditions in the Ottawa Valley is indicated by the rapid upward gradation from fresh-water to marine sediments (Gadd 1971).

It is suggested that the ice margin continued up the Bonnechere River valley toward White Partridge Lake (approximately 38 Km (24 M)) without a major change in orientation. At that time, it would appear that the ice margin had partly receded up the southward facing tilted fault block which forms the north side of the asymmetric Bonnechere River valley. As the ice margin receded upslope, melt-waters probably drained southward away from the ice and toward the Bonnechere River channel. An ice margin in that position (north of the White Partridge Lake sill) would have permitted

overflow of Lake Amable du Fond to join glacial melt-waters draining eastward through the Bonnechere River channel, which was confined between the ice margin and the north facing Bonnechere fault scarp (map 1).

Evidently, then, the Bonnechere sands were deposited as a proglacial delta, approximately 12,000 years ago, at the mouth of the Bonnechere River channel, where glacial melt-water runoff, augmented by glacial Lake Amable du Fond overflow, drained into Lake St. Lawrence (183 m (600 ft))(fig. 2 at Round Lake)(map 1).

The Bonnechere sand deposit (183 m (600 ft)) does not appear to have been entrenched by the Bonnechere River channel when Lake St. Lawrence (base level) drained to sea level (171 m (560 ft)). This indicates that the Bonnechere River channel was possibly underfit¹¹ prior to the drop of its base level (Lake St. Lawrence); and therefore, the Bonnechere River channel may have been abandoned as a principal route by proglacial melt-water runoff and lake overflow before the ice barrier was penetrated southeast of Quebec City.

It is further suggested that the ice margin ran west-northwest from near White Partridge Lake to Kilrush Lake, and from there north-northwest into the Mattawa River valley (map 1). The Rutherglen moraine (Harrison 1971, Chapman and Putman 1973) was possibly formed subaqueously in Lake Algonquin by still active ice at about that time. Boissonneau (1967) suggests that the ice margin may have run northward from the Mattawa Valley toward Balsam Creek, Ontario, and from there almost due west to the Whiskey Lake morainic belt. The close of the main Lake Algonquin stage was correlated by Boissonneau (1967) with:

- (1) the development of a reentrant in the ice margin in the area northeast and northwest of North Bay, Ontario, that was occupied during its development by successively lower stages of Lake Algonquin;
- (2) sand deposits in the reentrant area near Balsam Creek, whose elevations (375 m (1,230 ft)) "...indicate they were laid down during the initial

¹¹An underfit river is obviously too small for its present valley (Monkhouse 1970).

Fossmill outlet phase" (Boissonneau 1967): lake level was falling below its high stage 386 m (1,266 ft) relative to the isostatic rebounding land as the the reentrant proceeded northward. The above events are described in greater detail below in section 4.2d.

In summary, the distribution of morainic, deltaic, shoreline, and sea and lake bed deposits in the Bonnechere-Ottawa and Mattawa valleys indicates that the ice margin at the close of the main Lake Algonquin stage may have been, for the most part, in direct contact with glacial Lake Algonquin, Lake Amable du Fond, and Lake St. Lawrence between North Bay and Montreal (map 1).

Proglacial melt-water drainage augmented by overflow from Lake Amable du Fond (by way of the White Partridge Lake outlet) has been shown to be an important factor in the formation of the Bonnechere sand deposit. Drainage of proglacial melt-waters and lake overflow into the Bonnechere Valley, however, apparently ended before glacial Lake St. Lawrence drained to sea level, and marine conditions were initiated in the Ottawa and Bonnechere Valleys. The Bonnechere River Channel was apparently abandoned (as a principal proglacial drainage route) when a lower easterly route became available; a route which did not enter the upper Bonnechere Valley.

The related effects of ice recession, topography, and regional isostasy on glacial lake configuration and proglacial melt-waters and glacial lake overflow dispersal in the Ottawa-Bonnechere graben are analysed below.

4.1d Consideration of ice margin recession, Fossmill drainage, and sand deposition in the Ottawa-Bonnechere graben

Fossmill drainage period. The principles of isostatic uplift, assuming a constant rate of uplift over a short period of time, and an ice margin receding and halting at regular intervals, are illustrated in figure 25. This situation is possibly similar to that which existed during the closing

Figure 3

B Bernard Lake Beach, 1245 feet present elevation

SR South River Beach, 1220 feet present elevation

TC Trout Creek Beach, 1195 feet present elevation

Figure 25: Glacial isostasy and ice sheet recession model - volume of the proglacial lake is assumed to be constant through sequence A, B and C. During temporary halts in ice sheet recession shoreline features develop along a synchronous strandline. However, following further recession and isostatic rebound the shoreline emerges and shoreline features are raised above lake level. Due to the time transgressive advance of the shoreline to the north behind the receding ice front shoreline features that were formed further to the south and continued to rise with the rebounding crust tend to be at successively higher elevations. A line joining these features from south to north tends to decline in elevation at a more or less constant rate - it is referred to as a metachronous strandline. (after Flint 1971, fig. 13-11, p. 361)

stages of Lake Algonquin in the North Bay-Mattawa region. The volume of Lake Algonquin may have been extended over an increasing area as the ice margin and the ice marginal shoreline receded farther to the north. Consequently, assuming lake water volume remained about the same, lake level may have dropped (figure 25). At the same time the area just south of the ice margin was being isostatically uplifted rapidly relative to areas much farther to the south. Shorelines, therefore, were in a continuous state of emergence. Under those conditions synchronous development of shoreline features may have occurred for short periods of time at different elevations as the land was raised above water levels. The decline in elevation toward the north at a rate of 0.47 m/Km (2.5 ft/M) between the Lake Bernard, South River and Trout Creek beaches (Chapman 1954) is evidence of metachronous shoreline development¹² (fig 25).

The implications of a continually emerging Lake Algonquin shoreline on levels of Lake Amable du Fond is shown in figure 26. The highlands, on which are located the Kilrush Lake sill (348 m (1,140 ft) and Sobie-Guilmette Lakes sill (343 m (1,125 ft) (Harrison 1971), may have been undergoing isostatic uplift whilst the level of Lake Algonquin-Lake Amable du Fond water body (363 m (1,266 ft)) was falling and the ice margin was receding¹³ (fig. 26 and map 1, 3a-3c). Lake Algonquin and Lake Amable du Fond would have been separated when the highlands emerged (map 3a). Control of eastward draining overflow of Lake Algonquin could then have been transferred from the White Partridge Lake sill (381 m (1,250 ft)) to the Kilrush Lake sill (348 m (1,140 ft)) (map 1, 3a-3b). Other transfers of control over lake overflow could have occurred as the land progressively rose more

¹²Metachronous shoreline development involves shoreline features forming at different times and typically at successively lower elevations (different strand lines) as water levels fall through time.

¹³If one accepts that estimates of average rate of ice margin recession of between 290 m (950 ft) to 137 m (450 ft) per varve cycle (calculated by Hough (1956) and Antevs (1925), respectively, from varved clays in the North Bay-Timiskaming area) can be applied to the Lake Algonquin context, then the sequence of events shown in map 1, 3a-3c may have occurred within one hundred years.

(A) Correlates with
Map 1.

(B) Correlates with
transition between
Map 1 and 3a

(C) Correlates with
Map 3a and Map 3b

Figure 26: Idealized diagram showing three phases in the control of lake levels in the North Bay-Mattawa region. Deglaciated areas are subjected to submergence and uplift. Point X is the same in A, B, and C phases. At first the entire area is below lake level (A) (lake level is controlled by an ice dam across the Petawawa River basin near White Part-ridge Lake). Through uplift and extension of the lake northward the submerged Wistwasing-Amable du Fond drainage divide forms a shoal, but lake level on either side is the same (B). After further uplift and lake extension the level of Lake Algonquin falls below the elevation of the ridge. The shoal, as it emerged, became a controlling sill (first at Kilrush Lake, then, at Sobie and Guilmette Lakes) for lake overflow from Lake Amable du Fond (C). In this model the direction of drainage over the sills is opposite to that suggested by Harrison (1971).

(after Flint 1971, fig. 13-3, p. 350)

in elevation and sills at successively lower elevations emerged, i.e., Sobie-Guilmette Lakes sill (341 m) and Mink Lake sill (328 m) (map 3b and 3c).

The absence of crustacean populations in present lakes above 317 m (1,040 ft) and north of present Petawawa River is evidence that the north shoreline of Lake Amable du Fond (the receding ice margin) may have only been a short distance north of the present Petawawa River (between Lake Traverse and Koistkokwi Lake) before lake level fell from 386 m (1,266 ft) to approximately 317 m (1,040 ft). It is suspected therefore, that the ice dam persisted near Lake Amable du Fond as water levels continued to fall. If such was the case, the Kilrush Lake sill and Sobie-Guilmette sill, may have, in turn, controlled the level of Lake Amable du Fond. Therefore, Lake Amable du Fond may have drained both westward and eastward during its life span (westward flow in the upper reaches of the Wistiwasing and Nosbonsing Rivers, respectively) (fig. 24, fig. 25, map 3a and map 3b) (It should be noted here that the literature does not contain evidence to support the easterly direction of drainage through the Fossmill channels that has always been assumed).

Lake Amable du Fond would then drain by way of Greenleaf Creek and Carcajou Creek to the Pine River, and then southward to glacial Lake St. Lawrence when the Pine River sill (253 m (830 ft))¹⁴ was freed of ice (fig. 24, map 1). The ice sheet would have prevented drainage eastward over the lower and more northerly Barron River sill (224 m (800 ft))¹⁵ near Brawny, Ontario, and through present Grand Lake until further ice recession put the ice margin just north of the Kathmore sill (233 m (765 ft))¹⁶ of the Indian River channel (fig. 24, map 1).

No evidence has been found that would indicate Lake St. Lawrence (183 m (600 ft)) extended farther north than the Bonnechere Valley before

¹⁴ Sill elevation estimated from topographic map (NTS), 50 ft contour interval.

¹⁵ Sill elevation estimated from nearby benchmark identified on topographic map (NTS).

¹⁶ Sill elevation estimated from nearby benchmark identified on topographic map (NTS).

being drained to sea level. Burger (1967, p. 401) states:

In the valley of the Ottawa River....an ancient shoreline between 500 and 520 ft altitude probably marks the upper limit.... of the Champlain Sea.

Evidence of an ancient shoreline is based on a number of features, which form patterns, rather than on a series of related beaches. Between 500 and 520 ft, wave-cut terraces occur on high kames and eskers, whereas flat tops are a feature of lower kames. Bare bedrock occurs along the 520 ft contour, either as strips or as small rock 'islands' scattered along the border of areas of clay. Boulder concentrations (lag concentrations) occur at [520 ft]....Glaciofluvial deposits in the Ottawa Valley above approximately 520 ft usually have sharp features and distinct kettles; between 500 and 520 ft these deposits have rounded crests as if washed-over.[see fig.27].

The extensive deltaic deposits of stratified fine and medium sand near Petawawa have a maximum elevation of approximately 520 ft. They are considered ancient deltas formed in the Champlain Sea by the Ottawa, Petawawa, Barron, and Indian Rivers.... Areas of fine sand with nearly level topography occur between 490 ft and 520 ft. They are probably off-shore deposits, bordering on clay areas.

Marine fossils found near Pembroke (10,870 years B.P., GSC-90) have been attributed to the Champlain Sea. The above evidence indicates the Champlain Sea transgressed up the Ottawa Valley, north of the Bonnechere River valley, behind the receding ice margin.

Figure 27: View from the top of the 159 m (520 ft) shoreline scarp looking westward toward Ottawa River and highlands in Quebec. Photograph was taken about 50 m from kettle (159 + m) that has a distinct form (near Hiam, Ontario, map 1).

The lack of evidence for Lake St. Lawrence north of Bonnechere River suggests that the ice margin had not receded very far north from the position shown in map 1 before the lake drained to sea level. The fall in water level following the breaching of the ice barrier near Quebec City may have been, therefore, catastrophic. A rapid decline in base level (Lake St. Lawrence) is indicated by the present morphology of the courses established by the Pine River and Bonnechere River across the Bonnechere sand deposit (map 1). The Pine River channel entrenched the sand deposit along the deposit's northern perimeter. The directness of the route is evidence that little lateral displacement of the channel occurred while the channel was cut vertically. Consequently most of the sand deposit remained intact. The underfit Bonnechere River was incapable of eroding a channel through the emerged sand deposit. The river, instead, took a passive route northward and joined the Pine River channel above the knickpoint (map 1).

While the above events took place, water levels in the Huron basin may have continued to fall and the land rose by isostatic uplift. The Mink Lake sill (328 m (1,075 ft), Harrison 1971) (which is correlated with the Pine River and Indian River channels in maps 3b and 3c), therefore, possibly emerged shortly after its initiation. Emergence of the Mink Lake sill gradually reduced lake overflow discharge in the Fossmill system until the Mink Lake sill emerged and lake overflow into the Fossmill system ended. But, in considering the extensive catchment area of the Indian River at that time (map 1) it is clearly apparent that the loss of lake overflow into the Fossmill system would have reduced total discharge into the Champlain Sea at the mouth of the Indian River by an insignificant amount.

It would seem from the above that the Alice sand deposit (Indian River valley) was deposited into the Champlain Sea (159 m (520 ft)) by proglacial melt-waters and runoff from deglaciated areas, which were augmented by Fossmill lake overflow for part of the time of development of the Alice sand deposit.

Sedimentation of the Bonnechere sand deposit into Lake St. Lawrence was possibly initiated when proglacial melt-waters began to drain away from the receding ice margin by way of the Bonnechere River valley. At that time glacial Lakes Algonquin and Amable du Fond were possibly at their maximum level within the study area. Progressive decline in lake levels resulted after further recession of the ice margin northward of the Bonnechere River valley. The sequence of routes taken by eastward draining lake overflow from Lakes Algonquin and Amable du Fond was probably regulated, in part, by the effects of ice blockage of principal preglacial valleys. The topography of the east side of the Lake Huron basin is such that with the presence of the ice sheet no outlet channels of significant discharge capacity probably came into existence before lake levels fell below the elevation of the Mink Lake sill of the Fossmill outlet channel--the only potential route that is topographically in a position to permit significant easterly lake overflow (south of the Mattawa Valley). In all likelihood the supply of easterly lake overflow was terminated before sedimentation of the Alice sand deposit in the Indian River valley ended and Lake St. Lawrence drained to sea level. Evidently, the Bonnechere sand deposit was formed as a delta where proglacial melt-waters augmented by glacial lake overflow, draining first by way of the Bonnechere valley, then, the Pine River valley, entered Lake St. Lawrence.

Shortly after the Bonnechere and Pine River routes were abandoned by melt-waters and drainage by way of the Indian River valley was initiated Lake St. Lawrence possibly drained to sea level, thus initiating the Champlain Sea in the Ottawa-St. Lawrence lowlands. The Alice sand deposit was probably formed as a delta where Indian River melt-waters discharged into the Champlain Sea (marine limit).

Post-Fossmill drainage period. The many eskers in the study area which terminate at pre-glacial valleys which run approximately subparallel to the ice margin are evidence that as the ice margin receded further north of the Indian River glacial melt-waters drained into newly revealed pre-existing valleys. The topography is such, however, that no channel was in a position to capture catchment drainage from the Indian River channel until the ice margin cleared the Barron River as far east as the Champlain Sea. After the ice cleared the Barron River the present Indian River catchment was established east of the Kathmore sill and waters from the remaining catchment to the west of the sill began to drain eastward by way of the Barron River (map 1).

Sediments were deposited at the mouth of the Barron River in the Champlain Sea. This area is commonly referred to as the 'Petawawa sand plain region' (Chapman and Putman 1973) and is occupied by the Airport and Jorgens sand deposits. The sands in this region have been interpreted as deltaic (see General Statement above).

The Barron River continued to be the principal proglacial drainage route while ice recession continued because no further change in drainage pattern could occur until the ice, which blocked the Petawawa River north of present Lake Traverse, receded beyond the most northerly section of the Petawawa River; freeing drainage eastward through that channel to the Champlain Sea. Following that event, the Petawawa River captured all drainage originating west of the Barron River channel sill near Brawny (map 1). The present configurations of the Barron River and Petawawa River catchments were established as a result of these events: the Barron River presently joins Petawawa River, indicating that rerouting of water away from the Petawawa area did not occur thus the supply of sediments to the area was probably not reduced.

The ice margin, for a period of time during recession, encountered and ran approximately parallel to the Ottawa-Petawawa drainage divide. Many low places which, because of their size and topographic situation, seem most likely to have functioned as water gaps are indicated on map 1

(channels at Park Lake and Findley Lake were first suggested by Gadd (1963). Melt-waters originating from ice recession down the north facing flank of the divide, presumably were channelled through pre-existing drainage routes, namely present Moffat Creek, Chalk River (Gadd 1963), Wylie Creek (Gadd 1963), and the Ottawa River.

The only information available on sediments in the vicinities of the above channels is a surficial geology map by Gadd (1963). This map (in concurrence with topographic maps) indicates that (1) fluvial gravels, bed-rock, till and sand comprise much of the surface north of the Jorgens sand deposit; (2) most of the 'sand' covered surfaces are well above the marine limit (159 m (520 ft)). Since the development of sand deposits north of the Jorgens sand deposit is not a principal concern of the present study, no other information was collected on these sediments. However, it would seem reasonable to suggest that most of the sediments described above were deposited as till, or melt-water outwash and alluvium.

Summary. Presumably it took approximately 100 years (centered around 12,000 years ago) for the ice margin to recede from the White Part-ridge Lake sill to the Pine River sill, during which time it dammed Lake Amable du Fond. The Fossmill overflow system may have, therefore functioned for a period of approximately 100 years and not the 1000 year period suggested by Harrison (1971, table 4).

The principal factors that controlled dispersion of waters and sediment through the Fossmill system were the pre-existing fault-controlled topography and the receding ice margin.

The Bonnechere sands and Alice sands were deposited successively where proglacial melt-waters and Fossmill overflow entered glacial Lake St. Lawrence and the Champlain Sea, respectively.

Fossmill lake overflow was probably not a factor in the development of the Airport and Jorgens sand deposits, but rather these were deposited by proglacial melt-waters which drained through the Barron and Petawawa Rivers.

Following further ice sheet recession north of the Indian River basin melt-waters were probably redirected into the topographically lower Barron River valley; through which it flowed eastward to the Champlain Sea (marine limit) and formed the Airport sand deposit--an outwash delta. This delta continued to develop while the ice receded into the Petawawa River valley (lower part). That valley became the principal melt-waters drainage route.

4.2 Concluding Statement

The sequence of events leading to the development of the sand deposits described above is very much different from that generally interpreted in the literature, however, the end product--a series of deltaic sand deposits--is basically similar. The most significant differences in these interpretations are (1) the deltas are seen as developing time transgressively and not virtually contemporaneously as generally previously inferred in the literature, and (2) glacial lake overflow probably had a relatively small part in the sedimentation of the deltas and was probably not the principal component of discharge and sediment supply to the upper Ottawa Valley as it has generally been considered.

New evidence on sediment characteristics of sand deposits (Jorgens and Airport sands) in the 'Petawawa sand plains region' shows that only the Airport sand deposit had a deltaic genesis. The sandy plains northwest of C. F. B. Petawawa (Jorgens sand deposit) are probably not remnants of a deltaic depositional surface, but rather probably represent the formation of a beach-inshore sand complex on a shallow water bedrock platform along the shoreline of the Champlain Sea (marine limit). The interpretation of these sands (Jorgens sand deposit) as a littoral deposit is entirely a new concept of sedimentation processes in sand deposit formation in the upper Ottawa Valley.

Knowledge obtained from field and laboratory studies of the morphologic and intrinsic sedimentary characteristics of sand deposits in the upper Ottawa Valley has provided new insight into their history of sedimentation.

Strengthening of the concepts of sedimentation processes developed in this study will require further field analysis of paleogeographic elements and intrinsic and associated sedimentary characteristics of the sand deposits.

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APPENDIX

TRAND SURFACE ANALYSIS REPORT
 SURFACES FITTED TO THE 11
 LINE A - A' ON MAP 7

TABLE 1

PARTICLE SIZE ANALYSIS OF SEDIMENTS (PHI UNITS)

<u>Sample Number</u>	<u>Mean Size</u>	<u>Standard Deviation</u>	<u>Skewness</u>	<u>Kurtosis</u>
1	0.90	1.10	0.15	1.97
2	1.23	1.00	-0.41	2.41
3	1.06	1.09	-0.10	1.97
4	0.91	1.10	0.12	1.97
5	2.78	0.24	-4.15	29.22
6	2.41	0.47	-1.54	6.61
7	0.07	0.96	1.14	3.30
8	0.64	0.86	0.21	2.39
9	1.05	1.09	-0.07	1.98
10	0.85	0.96	0.02	2.24
11	1.83	0.64	-1.43	6.08
12	2.04	0.61	-0.76	3.68
13	2.05	0.75	-1.14	3.86
14	1.51	0.91	-0.65	2.61
15	0.87	1.06	0.22	2.06
16	2.04	0.81	-1.48	5.18
17	2.61	0.28	-1.38	6.96
18	2.76	0.22	-3.02	18.27
19	1.60	0.83	0.15	2.03
20	2.24	0.61	-1.25	4.76
21	1.42	0.93	-0.39	2.53
22	0.62	0.82	0.23	2.62
23	1.01	0.96	-0.14	2.13
24	0.93	0.79	0.08	2.78
25	1.22	0.91	-0.08	2.12
26	0.32	0.73	0.99	4.40
27	1.83	0.47	0.22	2.85
28	1.15	1.04	-0.28	2.20
29	1.10	1.13	-0.14	1.88
30	1.06	1.06	-0.13	2.05
31	1.00	1.10	0.01	1.96
32	1.54	0.39	0.24	3.95
33	2.28	0.48	-0.70	3.03

TREND SURFACE ANALYSIS RESULTS FOR 1ST, 2ND AND 3RD DEGREE TREND SURFACES FITTED TO THE 16 ELEVATION POINTS NORTH AND SOUTH OF LINE A - A' ON MAP 7 (JORGENS AND AIRPORT SAND DEPOSIT AREAS)

1st degree

NO.	ft Z-VALUE	ft VALUE	ft RESIDUAL	ERROR MEASURES
1	552.000	535.132	16.868	-----
2	523.000	519.373	3.627	
3	524.000	518.655	5.345	STANDARD DEVIATION 9.74
4	516.000	515.085	.915	
5	516.000	521.946	-5.946	VARIATION EXPLAINED
6	505.000	514.380	-9.380	BY SURFACE .21119497E 04
7	490.000	508.344	-18.344	
8	523.000	521.547	1.453	VARIATION NOT EXPLAINED
9	512.000	516.125	-4.125	BY SURFACE .14240503E 04
10	504.000	518.478	-14.478	
11	521.000	510.330	10.670	TOTAL VARIATION .35360000E 04
12	514.000	508.805	5.195	
13	506.000	502.220	3.780	COEFFICIENT OF
14	500.000	486.876	13.124	DETERMINATION .59727079
15	496.000	504.995	-8.995	
16	492.000	491.709	.291	COEFFICIENT OF CORRELATION .77283299

2nd degree

NO.	Z-VALUE	VALUE	RESIDUAL	ERROR MEASURES
1	552.000	550.008	1.992	-----
2	523.000	518.838	4.162	
3	524.000	517.575	6.425	STANDARD DEVIATION 7.87
4	516.000	511.779	4.221	
5	516.000	522.754	-6.754	VARIATION EXPLAINED
6	505.000	510.870	-5.870	BY SURFACE .26069561E 04
7	490.000	503.379	-13.379	
8	523.000	522.467	.533	VARIATION NOT EXPLAINED
9	512.000	513.459	-1.459	BY SURFACE .92904395E 03
10	504.000	515.980	-11.980	
11	521.000	506.089	14.911	TOTAL VARIATION .35360000E 04
12	514.000	503.944	10.056	
13	506.000	500.406	5.594	COEFFICIENT OF
14	500.000	496.382	3.618	DETERMINATION .73726130
15	496.000	500.823	-4.823	
16	492.000	499.248	-7.248	COEFFICIENT OF CORRELATION .85863918

3rd degree

NO.	Z-VALUE	VALUE	RESIDUAL	ERROR MEASURES
1	552.000	551.492	.508	-----
2	523.000	527.940	-4.940	
3	524.000	524.208	-.208	STANDARD DEVIATION 4.13
4	516.000	505.370	10.630	
5	516.000	514.534	1.466	VARIATION EXPLAINED
6	505.000	507.284	-2.284	BY SURFACE .32806934E 04
7	490.000	496.640	-6.640	
8	523.000	524.021	-1.021	VARIATION NOT EXPLAINED
9	512.000	512.176	-.176	BY SURFACE .25530655E 03
10	504.000	507.490	-3.490	
11	521.000	513.920	7.080	TOTAL VARIATION .35360000E 04
12	514.000	514.058	-.058	
13	506.000	507.543	-1.543	COEFFICIENT OF
14	500.000	499.550	.450	DETERMINATION .92779785
15	496.000	495.600	.400	
16	492.000	492.187	-.187	COEFFICIENT OF CORRELATION .96322256

Z-value = actual elevation value = calculated elevation

residual= (Z-value) - (value)

(Table 2 Continued On Following Page)

TABLE 2 CONTINUED

TREND SURFACE ANALYSIS RESULTS FOR 1ST, 2ND
AND 3RD DEGREE TREND SURFACES FITTED TO THE
10 POINT DATA GROUP NORTH OF LINE A - A' ON
MAP 7 (JORGENS SAND DEPOSIT AREA)

1st degree				2nd degree				3rd degree			
NO.	Z-VALUE	ft VALUE	ft RESIDUAL	NO.	Z-VALUE	ft VALUE	ft RESIDUAL	NO.	Z-VALUE	ft VALUE	ft RESIDUAL
1	552.000	551.456	.544	1	552.000	552.054	-.054	1	552.000	552.000	.000
2	523.000	523.072	-.072	2	523.000	523.405	-1.405	2	523.000	523.000	.000
3	524.000	521.705	2.295	3	524.000	523.403	.597	3	524.000	524.000	.000
4	537.000	536.525	.475	4	537.000	536.854	.146	4	537.000	537.000	.000
5	516.000	520.155	-4.155	5	516.000	517.568	-1.568	5	516.000	516.000	.000
6	505.000	506.503	-1.503	6	505.000	505.759	-.759	6	505.000	505.000	.000
7	490.000	489.423	.577	7	490.000	489.582	.018	7	490.000	490.000	.000
8	523.000	523.106	-.106	8	523.000	521.256	1.744	8	523.000	523.000	.000
9	512.000	511.588	.412	9	512.000	510.884	1.116	9	512.000	512.000	.000
10	504.000	502.468	1.532	10	504.000	503.836	.164	10	504.000	504.000	.000
ERROR MEASURES				ERROR MEASURES				ERROR MEASURES			
-----				-----				-----			
STANDARD DEVIATION				STANDARD DEVIATION				STANDARD DEVIATION			
1.04				1.77				.00			
VARIATION EXPLAINED				VARIATION EXPLAINED				VARIATION EXPLAINED			
BY SURFACE				BY SURFACE				BY SURFACE			
.27792966E 04				.27608206E 04				.27890000E 04			
VARIATION NOT EXPLAINED				VARIATION NOT EXPLAINED				VARIATION NOT EXPLAINED			
BY SURFACE				BY SURFACE				BY SURFACE			
.97033062E 01				.28179214E 02				.35762787E+06			
TOTAL VARIATION				TOTAL VARIATION				TOTAL VARIATION			
.27890000E 04				.27890000E 04				.27890000E 04			
COEFFICIENT OF DETERMINATION				COEFFICIENT OF DETERMINATION				COEFFICIENT OF DETERMINATION			
.99652082				.98989618				1.00000000			
COEFFICIENT OF CORRELATION				COEFFICIENT OF CORRELATION				COEFFICIENT OF CORRELATION			
.99825889				.99493521				1.00000000			

Z - value = actual elevation

value = calculated elevation

residual = (Z - value) - (calculated value)

(TABLE 2 CONTINUED ON FOLLOWING PAGE)

TABLE 2 CONTINUED

TREND SURFACE ANALYSIS RESULTS FOR 1ST, 2ND
AND 3RD DEGREE TREND SURFACES FITTED TO THE
6 POINT DATA GROUP SOUTH OF LINE A - A' ON
MAP 7 (AIRPORT SAND DEPOSIT AREA)

1st degree				2nd degree				3rd degree			
NO.	Z-VALUE	VALUE	RESIDUAL	NO.	Z-VALUE	VALUE	RESIDUAL	NO.	Z-VALUE	VALUE	RESIDUAL
11	521.000	523.144	-2.144	11	521.000	521.000	.000	11	521.000	521.000	.000
12	514.000	510.690	3.310	12	514.000	514.000	.000	12	514.000	514.000	.000
13	500.000	497.766	2.234	13	500.000	500.000	.000	13	506.000	506.000	.000
14	506.000	504.121	1.879	14	506.000	506.000	.000	14	500.000	500.000	.000
15	496.000	499.670	-3.670	15	496.000	496.000	.000	15	496.000	496.000	.000
16	492.000	493.608	-1.608	16	492.000	492.000	.000	16	492.000	492.000	.000
ERROR MEASURES				ERROR MEASURES				ERROR MEASURES			
-----				-----				-----			
STANDARD DEVIATION				STANDARD DEVIATION				STANDARD DEVIATION			
2.83				.00				.00			
VARIATION EXPLAINED				VARIATION EXPLAINED				VARIATION EXPLAINED			
BY SURFACE				BY SURFACE				BY SURFACE			
.57287329E 03				.61300000E 03				.61300000E 03			
VARIATION NOT EXPLAINED				VARIATION NOT EXPLAINED				VARIATION NOT EXPLAINED			
BY SURFACE				BY SURFACE				BY SURFACE			
.40126666E 02				.59604645E-07				.12516975E-05			
TOTAL VARIATION				TOTAL VARIATION				TOTAL VARIATION			
.61300000E 03				.61300000E 03				.61300000E 03			
COEFFICIENT OF DETERMINATION				COEFFICIENT OF DETERMINATION				COEFFICIENT OF DETERMINATION			
.93454032				1.00000000				1.00000000			
COEFFICIENT OF CORRELATION				COEFFICIENT OF CORRELATION				COEFFICIENT OF CORRELATION			
.96671623				1.00000000				1.00000000			

Z - value = actual elevation

value = calculated elevation

residual = (Z - value) - (calculated value)

```

1, C MECHANICAL ANALYSIS OF SEDIMENTS BY MOMENTS-HOOPER, GEOLOGY
2, C
3, C STAT IS SAMPLE NUMBER
4, C X IS CLASS MIDPOINT, F1 IS RAW FREQ, F IS PERCENT
5, C DIMENSION X(16),F(16),F1(16),ARRAY(150,16),FSUM(150),KSTAT(150)
6, C X FORMAT IS 103, FREQ AND STAT FORMAT IS 102
7, C 2 FORMAT(7X,13,2X,4F11.2)
8, C N=0
9, C NCT=0
10, C CALL HEAD1
11, C DO 707 IJ=1,10
12, C 707 FSUM(IJ)=0.0
13, C 102 FORMAT(16F5.2/13)
14, C 103 FORMAT(16F5.3)
15, C READ 103,(X(I),I=1,16)
16, C DO 10 K=1,10
17, C 11 READ 102,(F1(I),I=1,16),STAT
18, C N=N+1
19, C KSTAT(N)=STAT
20, C DO 909 II=1,16
21, C 909 FSUM(N)=FSUM(N)+F1(II)
22, C DO 910 JJ=1,16
23, C 910 F(JJ)=(F1(JJ)/FSUM(N))*100.0
24, C 41 SUMP = 0.0
25, C DO 20 I=1,16
26, C 20 SUMP = SUMP + X(I)*F(I)
27, C SUMP IS SIGMA FX
28, C SUMFR = 0.0
29, C DO 40 I=1,16
30, C 40 SUMFR = SUMFR + F(I)
31, C SUMFR IS SIGMA F. IDEALLY 100 PERCENT
32, C AVX = SUMP/SUMFR
33, C SVX IS ARITHMETIC MEAN IS SIGMA FX DIVIDED BY SIGMA F
34, C AVX IS FIRST MOMENT
35, C SUMPS = 0.0
36, C DO 60 I=1,16
37, C 60 SUMPS = SUMPS+F(I)*(X(I)-AVX)**2
38, C SUMPS IS SIGMA F(X-MEAN)SQUARED
39, C SMOM=SUMPS/SUMFR
40, C SMOM IS SECOND MOMENT
41, C SUMPT = 0.0
42, C DO 80 I=1,16
43, C 80 SUMPT = SUMPT+F(I)*(X(I)-AVX)**3
44, C SUMPT IS SIGMA F(X-MEAN)CUBED
45, C TMOM = SUMPT/SUMFR
46, C TMOM IS THIRD MOMENT
47, C SUMPPF = 0.0
48, C DO 100 I=1,16
49, C 100 SUMPPF = SUMPPF+F(I)*(X(I)-AVX)**4
50, C SUMPPF IS SIGMA F(X-MEAN) TO FOURTH POWER
51, C FMOM = SUMPPF/SUMFR
52, C FMOM IS FOURTH MOMENT
53, C STDEV = SQR(F(SMOM))
54, C SKEW = TMOM/STDEV**3
55, C CURT = FMOM/STDEV**4
56, C STANDARD DEVIATION, SKEWNESS AND KURTOSIS
57, C IF(SKEW)12,13,13
58, C 12 SKEW=SKEW-0.005
59, C GO TO 14
60, C 13 SKEW=SKEW+0.005
61, C 14 AVX=AVX+0.005
62, C STDEV=STDEV+0.005
63, C CURT=CURT+0.005
64, C IF(NCT.GE.32)CALL HEAD1;NCT=0
65, C PRINT 2,KSTAT(N),AVX,STDEV,SKEW,CURT
66, C NCT=NCT+1
67, C DO 928 I=1,10
68, C 928 ARRAY(K,I)=F(I)
69, C 10 CONTINUE
70, C CALL HEAD2(X)
71, C NCT=0
72, C DO 808 JK=1,10
73, C 808 FORMAT(13(F6.3)/3(F6.3),13)
74, C PRINT 809,KSTAT(JK),ARRAY(JK,12),L2=1,16),FSUM(JK)
75, C NCT=NCT+1
76, C 809 FORMAT(4X,13,2X,16(F6.3,1X),F6.2)
77, C IF (NCT.GE.38) CALL HEAD2(X); NCT=0
78, C PRINT 601
79, C 601 FORMAT ('1')
80, C 160 CALL EXIT
81, C END

```

(after Nickling 1972)

Table 4

HYDRAULIC GEOMETRY OF STREAM CHANNELS

$$\text{Mean flow velocity (v)} = \frac{d^{2/3} s^{1/2}}{n}$$

$$\begin{aligned} \text{Sediment transport concentration (C)} \\ = \frac{(vd)^{0.5} s^{1.5}}{n^4} \end{aligned}$$

where d is mean flow depth
 s is channel bed slope
 n is flow resistance

(after Leopold, Wolman, and Miller 1964)

SYMAP POLYNOMIAL TREND FUNCTIONS

In the statistical analysis by SYMAP data point elevation, variable Z , was the dependent variable and space coordinates (X and Y) the independent variables in polynomial trend functions fitted by least squares. The equations for a first-degree, three variate trend surface is,

$Z_{\text{trend}} = A + BX + CY$, and the equation for a second-degree, three variate trend surface is,

$$Z_{\text{trend}} = A + BX + CY + DX^2 + EXY + FY^2$$

The coefficients A , B , C , D , E , and F in the above equations are computed so that the sum of the squared deviations is a minimum (Harbaugh and Merriam 1968). Any number of terms can be added to the general polynomial equation to give higher degree trend surfaces.

Trend surfaces with increasing numbers of terms were fitted until the improvement in per cent of total sum of squares (R^2) became negligible, i.e., approximately 0.1 per cent.

SYMAP 'JOB SETUP' FORTRAN IV

Each line below is punched on one computer card. For data input format see SYMAP USER'S REFERENCE MANUAL, CARLETON UNIVERSITY EDITION.

```
! JOB (ACCT#,ACCT NAME) (CLASS 8)
! LIMIT
! ASSIGN F:6,(DEVICE,L2)
! LDEV L2,(FORM,SYM8)
! RUN (LMN,SYMAP,SYMAP)
!DATA
```

PREVU 'JOB SETUP' FORTRAN IV

There is no published reference for PREVU at present. Information on PREVU can be obtained from COMPUTING SERVICES, CARLETON UNIVERSITY.

```
! JOB (ACCT#,ACCT NAME) (CLASS 8)
! LIMIT (TIME,1),(LO,100),(CORE,24)
! ASSIGN F:8,(FILE,SVUDATA),(IN),(SAVE)
! RUN (LMN,PREVU,SYMAP)
! DATA
TYPE =
AZ =
*PLOT
!EOD
```

Elective 38. - Trend Surface Analysis
(1 card)

Card 1: Col. 4-5 '38' to identify the elective.
 Col. 11-20 Any number from 1 through 6 if the trend surface is to be calculated. The number gives the desired order of the trend surface. A number from 11 to 16 if a residual surface is to be calculated. The last digit indicates the desired order of the trend surface upon which the residual surface is to be based. All numbers must end in Col. 20 or have a decimal point.

This elective provides the user with an alternative method of producing contour maps, trend surface analysis. When a contour map is desired and Elective 38 is specified, SYMAP will not employ its usual interpolation procedure to determine values at each point location. Instead, the parameters for an equation representing a surface will be estimated, using a least-squares criterion. That is, the surface will be fitted to the data values in such a way that the sum of the squared deviations between the given values at data points and the height (value) of the computed surface at those data points is minimized.

Unlike SYMAP's standard interpolation procedure, the method of trend surface does not fit a surface so that it passes through each data point value given in the E-VALUES package. Although a perfect fit is possible, it is unlikely to occur in most applications. Therefore, there will tend to be residual values (the deviation, or difference, between the interpolated surface and the trend surface), which indicate local variations not predicted by the general trend represented by the fitted surface. Residuals may represent systematic local aberrations and they may be statistically explained by the incorporation of additional variables, as in multiple linear regression, an allied technique, or the residual values may simply exist due to measurement errors or other random noise. SYMAP will map the residual values instead of the computed trend values according to the user's wish.

As mentioned above, a trend surface is a statistically derived equation to explain variations in given data values (z-values) distributed either regularly or irregularly in x-y space. The equation describing the trend surface can be linear, in which case the surface will be a plane; quadratic (having a squared term), in which case the surface will be a paraboloid; cubic (having a cubed as well as squared and linear terms), giving an additional point of inflection; and

(from SYMAP User's Reference Manual, Carleton University Edition, Nov. 1976)

so on, up to the sixth degree for an equation with sixth order terms and lower order terms plus a constant, with the highest terms taken to the sixth power.

The higher the order of the surface, the more the residuals will be minimized and the more computation will be required. Users may request, then, a trend surface of any order from 1 to 6. Higher-order surfaces will reflect the variation in z-values more accurately, but lower-order surfaces may be quite useful in isolating important local trends from those that exist over a larger area. Trend surface analysis can therefore be considered as a process of filtering an input signal, the z-values, where the surface represents the resultant signal after filtering. The order of the surface determines the upper limit of variability, or frequency, of the input data which will pass through the filter; localized variation, or noise, will be blocked by the filter when lower orders are used, and it will be increasingly transmitted as the order of the surface increases. This is in contrast to standard SYMAP interpolation in which all data values are taken to be of equal significance in computing a contour surface.

If any of the Electives 31 through 36 or D-BARRIERS are used, they will not affect the normal calculations of a trend surface map since the surface is a fitted one and not interpolated. However, they will affect a map of residuals because the interpolated surface, from which the trend values are subtracted, will be affected.

For some sources on various applications, see:

Chorley, R.J. and Haggett, P.
 "Trend-Surface Mapping in Geographical Research",
Transactions, Institute of British Geographers, No. 37,
 1965, pp. 47-67.

Harbaugh, John W. and Merriam, Daniel F.
 "Computer Applications in Stratigraphic Analysis",
 New York, Wiley, 1967, Chapter 5.

Default: Trend surface analysis will not be used.

TABLE 5

GENETIC SORTING CLASSIFICATION BASED ON STANDARD DEVIATION

Sorting Interval	Sorting Designation	Environment of Deposition
Medium to fine and very fine-grained sands (mean $>1.0-2.0\phi$):		
<0.35	Very well sorted	Most coastal, barrier bar, and lake dune sands, many beach sands, many marine sands above wave base, many lagoonal sands
$0.35-0.50$	Well sorted	Most beach sands, many or most marine sands above wave base, many lagoonal sands, many inland dune sands, some river sands
$0.50-0.80$	Moderately well sorted	Most river sands, many beach sands (0.80 is approx. upper limit for beach sands), many lagoonal sands from restricted lagoons, most continental-shelf sands below wave base, most inland dune sands (0.80 is approx. upper limit except for some stable dunes)
$0.80-1.40$	Moderately sorted	Many river sands (1.40 is approx. upper limit for river sands) some lagoonal sands from restricted lagoons, some continental shelf sands below wave base, many glaciofluvial sands
$1.40-2.00$	Poorly sorted	Many glaciofluvial sands
$2.00-2.60$	Very poorly sorted	Many glaciofluvial sands
>2.60	Extremely poorly sorted	Some glaciofluvial sands
Coarse-grained sands (mean $<1.00\phi$):		
$0.50-0.80$	Moderately well sorted	Many beach sands
$0.80-1.40$	Moderately sorted	Most river sands, many or most beach sands, most continental-shelf sands
$1.40-2.00$	Poorly sorted	Some river sands, some continental-shelf sands, many glaciofluvial sands
$2.00-2.60$	Very poorly sorted	Many glaciofluvial sands
>2.60	Extremely poorly sorted	Some glaciofluvial sands

(from Friedman 1962)

SEDIMENT SAMPLING AND GRAIN-SIZE MEASUREMENT PROCEDURES

Field Procedures. A Soil Test core auger sampler with a 10 cm barrel was used to collect core samples at a depth of approximately 1 m. Grab samples from sections at gravel pits were taken from the pit face using a 15 cm trowel.

Due to time limitations sample sites were selected on the basis that sediments collected from those sites appeared to be representative of local sedimentary characteristics. The sampling plan was meant to obtain information on sediment grain-size characteristics which could be related to the local morphologic characteristics of the sand deposits.

Laboratory Procedures. Samples were split to 100 gm size, then sieved 20 minutes with a sieve shaker using sieve sizes between -1.25 ϕ (mesh no. 8, Tyler Sieves, 2.58 mm) and +3.00 ϕ (mesh no. 120, Tyler Sieves, 0.125 mm). Selected samples from along the Brindle Road traverse were analysed between sieve sizes -2.00 ϕ and +4.25 ϕ (see figs. 13 and 15) so that to analyse the distribution of fines; however it was considered unnecessary to analyse all 33 samples within these size limits because of the general lack of fines (less than 10 per cent), i.e., grain-sizes less than +3.00 ϕ . Sediments that collected on individual sieves were weighted to the nearest 0.01 gm using a Mettler scale. This raw weight data was used to compute moment measure parameters (see computer program listing, Appendix, p.).

SAND DEPOSITS AND HYDROLOGY

MAP 1 MAIN LAKE ALGONQUIN-LAKE ANABLE du FOND WATER BODY
386 m (1266 ft) with POSITION OF THE ICE MARGIN
APPROXIMATELY 12,000 YEARS AGO and GLACIAL LAKE
ST. LAWRENCE 183 m (600 ft)

LEGEND

1 BONNECHERE SAND DEPOSIT	▲ MELT-WATER CHANNEL SILL	--- POSSIBLE ICE MARGIN	D BERNARD LAKE BEACH --LAKE ALGONQUIN
2 ALICE SAND DEPOSIT	● CRUSTACEAN POPULATION WITHIN BOUNDARIES OF GLACIAL LAKES	--- DRAINAGE DIVIDE	E SOUTH RIVER BEACH --LAKE ALGONQUIN
3 AIRPORT SAND DEPOSIT	■ TOWN SITE	--- OBSOLETE STREAM ON FAULT SCARP	F TROUT CREEK BEACH --LAKE ALGONQUIN
4 JORGENS SAND DEPOSIT	→ PROGLACIAL DRAINAGE ROUTE	B WATERGAP	G MINK LAKE SILL 323 m (1075 ft)
5 PEMBROKE SAND DEPOSIT	↘ PROGLACIAL WATERGAP	C WATERGAP	H KOISTOKWI LAKE
○ GLACIAL LAKE OUTLET SILL	--- POSSIBLE GLACIAL LAKE SHORELINE	I MINK LAKE MORAINES	

1:460,000

SAND DEPOSITS AND HYDROLOGY

MAP 3a GLACIAL LAKE OUTLET AT KILRUSH LAKE 348 m (1140 ft)
POSSIBLE WESTERLY OVERFLOW OF GLACIAL LAKE AMABLE DU FOND

LEGEND

1. BONNECHERE SAND DEPOSIT	▲ MELT-WATER CHANNEL SILL	--- POSSIBLE ICE MARGIN	D. BERNARD LAKE BEACH
2. ALICE SAND DEPOSIT	● CRUSTACEAN POPULATION WITHIN BOUNDARIES OF GLACIAL LAKES	--- DRAINAGE DIVIDE	E. SOUTH RIVER BEACH
3. AIRPORT SAND DEPOSIT	■ TOWN SITE	--- OBSOLETE STREAM ON FAULT SCARP	F. TROUT CREEK BEACH
4. JORGENS SAND DEPOSIT	→ PROGLACIAL DRAINAGE ROUTE	B. WATERGAP	G. MINK LAKE STILL
5. PEMBROKE SAND DEPOSIT	↘ PROGLACIAL WATERGAP	C. WATERGAP	H. KOISTOKWNI LAKE
○ GLACIAL LAKE OUTLET SILL	--- POSSIBLE GLACIAL LAKE SHORELINE		I. MINK LAKE MORAINES

1:460,000

■ BALSAM CREEK

SAND DEPOSITS AND HYDROLOGY

N

MAP 3b GLACIAL LAKE OUTLET AT SOBIE-GUILMETTE LAKES 343 m (1125 ft)
POSSIBLE WESTERLY OVERFLOW OF GLACIAL LAKE AMABLE DU FOND

LEGEND

1 ■ BONNECHERE SAND DEPOSIT	▲ MELT-WATER CHANNEL SILL	--- POSSIBLE ICE MARGIN	D ■ BERNARD LAKE BEACH --LAKE ALGONQUIN
2 ■ ALICE SAND DEPOSIT	● CRUSTACEAN POPULATION WITHIN BOUNDARIES OF GLACIAL LAKES	--- DRAINAGE DIVIDE	E ■ SOUTH RIVER BEACH --LAKE ALGONQUIN
3 ■ AIRPORT SAND DEPOSIT	■ TOWN SITE	--- OBSEQUENT STREAM ON FAULT SCARP	F ■ TROUT CREEK BEACH --LAKE ALGONQUIN
4 ■ JORGENS SAND DEPOSIT	▲ PROGLACIAL DRAINAGE ROUTE	B ■ WATERGAP	G ■ MINK LAKE SILL 323 m (1075 ft)
5 ■ PEMBROKE SAND DEPOSIT	▲ PROGLACIAL WATERGAP	C ■ WATERGAP	H ■ KOISTOKWIT LAKE
● GLACIAL LAKE OUTLET SILL	--- POSSIBLE GLACIAL LAKE SHORELINE	I ■ MINK LAKE MORAINES	

1:460,000

SAND DEPOSITS AND HYDROLOGY